THE MESOPROTEROZOIC BELT SUPERGROUP

Jeffrey D. Lonn,[1] Russell F. Burmester,[2] Reed S. Lewis,[3] and Mark D. McFaddan[4]

[1]*Montana Bureau of Mines and Geology, Butte, Montana*
[2]*Western Washington University, Bellingham, Washington*
[3]*Idaho Geological Survey, Moscow, Idaho*
[4]*North Idaho College, Couer D'Alene, Idaho (retired)*

INTRODUCTION

The Belt Supergroup of western Montana, central and northern Idaho, and southeastern British Columbia, where it is called the Purcell Supergroup, is one of the largest and best-studied Mesoproterozoic sedimentary basins in the world (fig. 1). Even so, its immense thickness of more than 15 km, enormous extent covering >200,000 km^2, repetitious and monotonous lithologies, dismemberment by multiple tectonic events, and obscuration by magmatism and metamorphism have confounded geologists for more than a century. Its lack of fossils makes facies interpretations uncertain, but the absence of burrowing animals also allowed exquisite preservation of sedimentary features that have inspired study and offer important insights into the vast, barren Mesoproterozoic landscapes that are completely alien to our modern world.

Unusual characteristics of the Belt that have been difficult to explain include the succession's immense thickness without a single documented unconformity or marine sequence boundary, the enormous lateral extent of lithofacies with only very gradual facies changes, and the repetition of lithologies that reflect mainly shallow water and subaerial deposition. Winston and Link (1993) proposed that the complete absence of plants on Mesoproterozoic Earth created titanic, desolate landscapes that are completely different from any modern environments. Perhaps that is what is most interesting about the Belt: what it can tell us about these alien landscapes, the vast shallow lakes and seas, and the tectonic forces at work on this part of Mesoproterozoic Earth.

The "Belt" name comes from the Belt Mountains near Helena, which in turn derived their name from Belt Butte. Ironically, Belt Butte's belt is a Cretaceous sandstone bed. Belt rocks were recognized late in the 19th century and, because they host important mineral deposits, scientific study supported by the USGS began at the turn of the century. Belt rocks were initially studied in widely separated areas, each with its own formation names. Walcott (1899), Weed (1899), and Barrell (1906) worked around Helena, the Big Belt Mountains, and the Little Belt Mountains. Willis (1902) worked in Glacier National Park, Ransome and Calkins (1908) studied the Coeur d'Alene Mining District, and Calkins and Emmons (1915) mapped the Philipsburg area. Umpleby (1913) recognized similar Proterozoic sedimentary rocks in east-central Idaho. As work proceeded outward from these areas, correlations were suggested and then repeatedly modified, and depositional environments were proposed and argued over. Important to the understanding of the Belt Basin were a series of regional-scale USGS maps that cover almost the entire basin (Griggs, 1973; Miller and Yates, 1976; Mudge and others, 1982; Ruppel and others, 1993; Harrison and others, 1986, 1992; Wallace and others, 1986; Reynolds and Brandt, 2006a,b, 2007) and provided a solid foundation for subsequent work. In 1993, Winston and Link published a comprehensive review of the Belt Supergroup, citing 232 references up to that point! We will not repeat their work in this paper, but instead will review post-1993 publications to revise the concepts presented in that review.

Since their 1993 summary, Don Winston and Paul Link, with their students, have continued their tireless efforts to understand Belt sedimentology and its implications for depositional environments. John Lydon and Trygve Höy made especially important contributions from Canada. When the USGS abandoned the geologic mapping program that had provided so much information on the Belt, the Idaho and Montana geological surveys and universities took over, and have since completed many more detailed maps within the Belt Basin. Most recently, U-Pb dating of detrital zircons in Belt strata has helped generate, confirm,

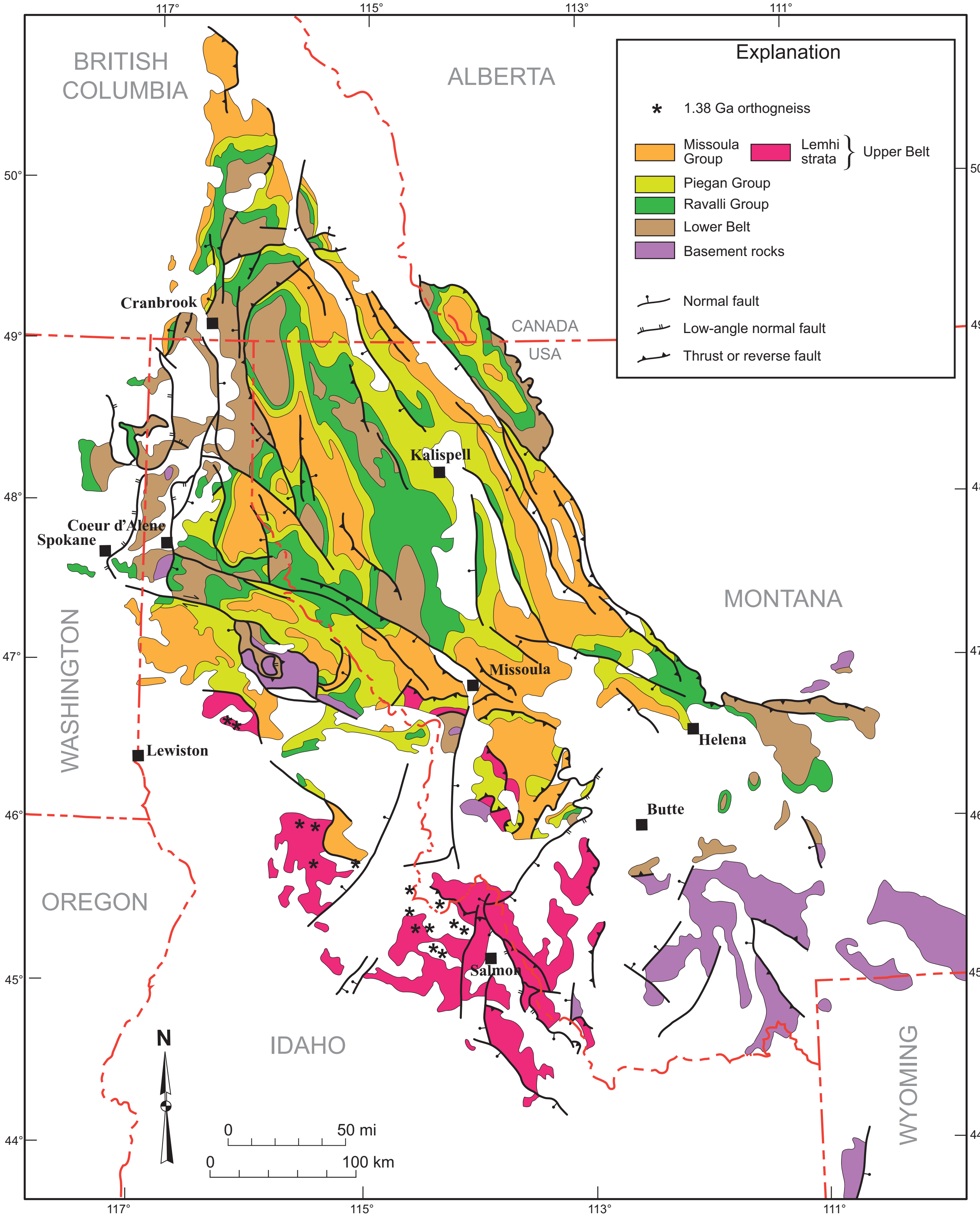


Figure 1. The Belt Supergroup is divided into the five Groups shown here. Note that the Missoula Group and Lemhi strata (upper Belt) are at least partly contemporaneous, and the contact between them is a fuzzy one marking gradual facies changes. Map is compiled and modified from Höy and others (1995), Vuke and others (2006), and Lewis and others (2012).

or reject many hypotheses. Although many questions remain—and Belt workers would not want it any other way—great progress has been made in understanding correlations, facies changes, and the younger tectonic features that isolate Belt exposures.

Most workers favor the interpretation that the Belt Basin was intracratonic, formed within the Columbia-Nuna supercontinent by rifting (Winston and Link, 1993; Ryan and Buckley, 1998; Sears and others, 1998; Lydon, 2005) or collision (Ross and Villenueve, 2003). Most also agree that the Lower Belt strata formed in a marine or restricted marine depositional environment, but many still argue about the overlying strata that were mostly deposited in shallower water. Some (Winston and Link, 1993; Lyons and others, 1998) proposed a lacustrine environment, whereas others (Wallace, 1998; Scheiber, 1998; Tysdal 2000a, 2003; Pratt, 2001, 2017a,b; Johnson, 2013) contended that the basin was open to the ocean and therefore marine for most of its history. In this paper, the use of the term “Belt Sea” is not meant to favor its interpretation as marine; the word “sea” can also be used to describe a large lake such as the Sea of Galilee (freshwater with an outlet) or the Dead Sea (salt water with no outlet). No disrespect to lacustrine proponents is intended.

REGIONAL BELT STRATIGRAPHY AND CHRONOLOGY

We organize our summary of prior work in the Belt Supergroup by Group, and include discussion of strata that are not technically included in the Groups but that have been correlated with them. Table 1 shows postulated correlations across the Belt Basin. Figure 2 illustrates the cumulative thickness of Belt strata shown in table 1 and the limited age constraints on the timing of deposition. These U-Pb geochronology data suggest that most of the strata accumulated rapidly between about 1470 and 1380 Ma (Winston and Link, 1993; Evans and others, 2000; Ross and Villeneuve, 2003; Anderson and Davis, 1995; Evans and others, 2000). Paleomagnetic data (Elston and others, 2002) from rocks that overlie the lower Belt are consistent with these dates. Interestingly, after more than 15 km (9 mi) of sediment accumulated over a roughly 100-million-year time span, disconformities and subtle unconformities with most overlying Phanerozoic strata indicate that much of the Belt remained largely undisturbed until the Sevier-Laramide orogeny 1.2 billion years later!

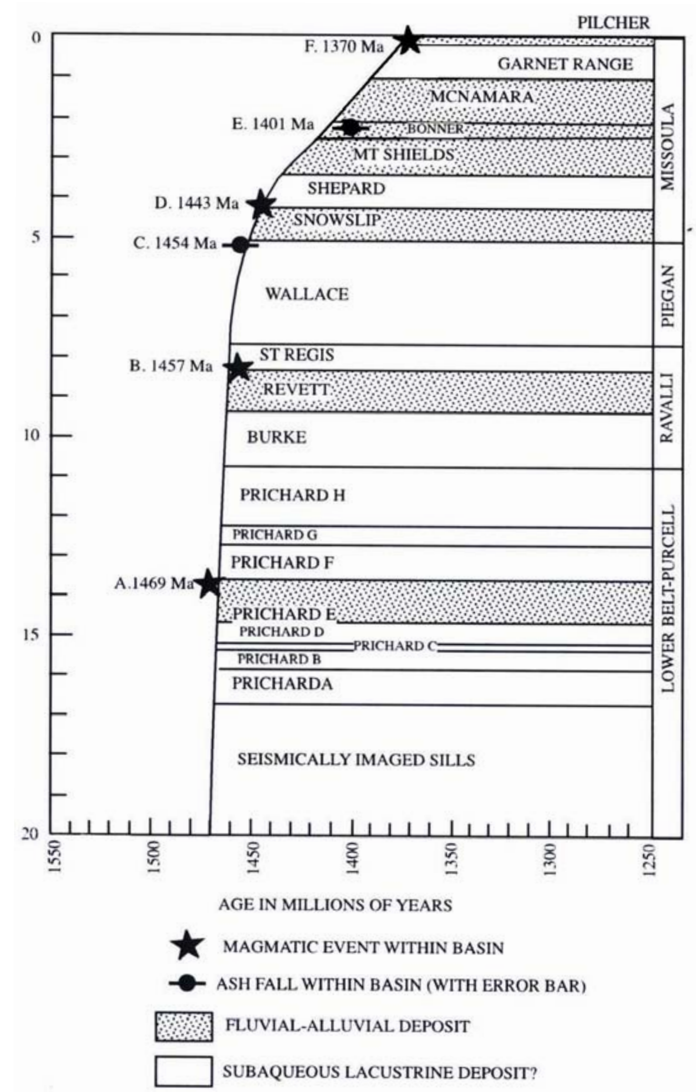


Figure 2. Sediment-accumulation curve for the Belt Supergroup in the central Belt–Purcell basin in western Montana is based on stratigraphic thicknesses from Winston (1986) and Cressman (1989), and U-Pb dates from (A) Anderson and Davis (1995), Sears and others (1998); (B) Sears and others (1998); (C,D,E) Evans and others, 2000; (F) Doughty and Chamberlain (1996). Stars indicate rifting events with mafic magmatism. Circles are dates from ash beds within the basin. Modified from Sears (2007a). Note that Constenius and others (2017) recently obtained significantly younger ages for D of 1338–1386 Ma.

Figures 3 and 4 are fence diagrams constructed across the basin showing lithologies, facies changes, and correlations. In the central part of the basin, the Belt Supergroup is divided into four groups: the lower Belt, Ravalli Group, Piegan Group, and Missoula Group (figs. 3, 4; table 1). Mesoproterozoic strata in the Lemhi Subbasin of east-central Idaho (Burmester and others, 2016) are at least partly correlative with the Missoula Group of Montana (table 1). The Deer Trail Group of northeastern Washington is also considered partly correlative with the Belt (Winston and Link, 1993), but 1100 Ma detrital zircons in the upper

Table 1. Belt-Purcell Supergroup Correlations

	Western Montana-Northern Idaho Winston & Link, 1993	Glacier National Park (NE Belt basin) Winston & Link, 1993	Helena Embayment (SE Belt basin) Winston & Link, 1993	British Columbia-Alberta Lydon, 2007	East Central Idaho Burmester and others, 2015, 2016 Lonn, 2017	East Central Idaho Winston and others, 1999	Eastern Washington Winston and Link, 1999 Farooqui, 1998
Missoula Group	Pilcher Fm				Apple Creek Fm		Deer Trail Group: Buffalo Hump Fm ?
Missoula Group	Garnet Range Fm; Libby Fm	Garnet Range Fm		Rooseville Fm			
Missoula Group	McNamara Fm; Libby Fm	McNamara Fm			Lawson Creek Fm	Lawson Creek Fm	
Missoula Group	Bonner Fm	Bonner Fm		Phillips Fm	Swauger Fm; Hoodoo Fm	Swauger Fm	
Missoula Group	Mt Shields Fm	Mt Shields Fm		Gateway Fm	Lemhi Group: Gunsight Fm; Yellow-jacket Fm	Lemhi Group: Gunsight Fm	Stensgar Dolomite
Missoula Group	Shepard Fm; Upper Wallace Fm	Shepard Fm		Sheppard Fm	Lemhi Group: Yellow Lake Fm; West Fk; Inyo Cr		McHale Slate
Missoula Group	Snowslip Fm; Upper Wallace Fm	Snowslip Fm		Nicol Creek Fm	Lemhi Group: Big Creek Fm		Edna Dolomite
Piegan Group	Wallace Fm	Helena Fm		Van Creek Fm		Lemhi Group: Yellow Lake Fm; Apple Creek Fm	Togo Fm
Piegan Group	Helena Fm		Helena Fm	Kitchner Fm			
Ravalli Gp	St Regis Fm; Empire Fm	Empire Fm	Empire Fm	Creston Fm		Lemhi Group: Big Creek Fm; Hoodoo Fm	
Ravalli Gp	Revett Fm	Spokane Fm	Spokane Fm				
Ravalli Gp	Burke Fm					Lemhi Group: West Fk Fm; Yellow-jacket Fm	
Lower Belt	Prichard Transition	Appekunney Fm	Greyson Fm	Aldridge Fm		Lemhi Group: Inyo Cr Fm	
Lower Belt	Prichard Fm	Altyn Fm	Newland Fm				
Lower Belt			Chamberlain Fm				
pre-Belt	Gold Cup quartzite; Marble Creek quartzite		Neihart Fm	Ft. Steele Fm			

Note. Belt Supergroup correlations were compiled and modified from Winston and Link (1993), Farooqui (1998), Winston and others (1999), Lydon (2007), Burmester and others (2015, 2016), and Lonn (2017). Note the contrasting correlations in east-central Idaho and the uncertainty in the eastern Washington Deer Trail Group correlatives.

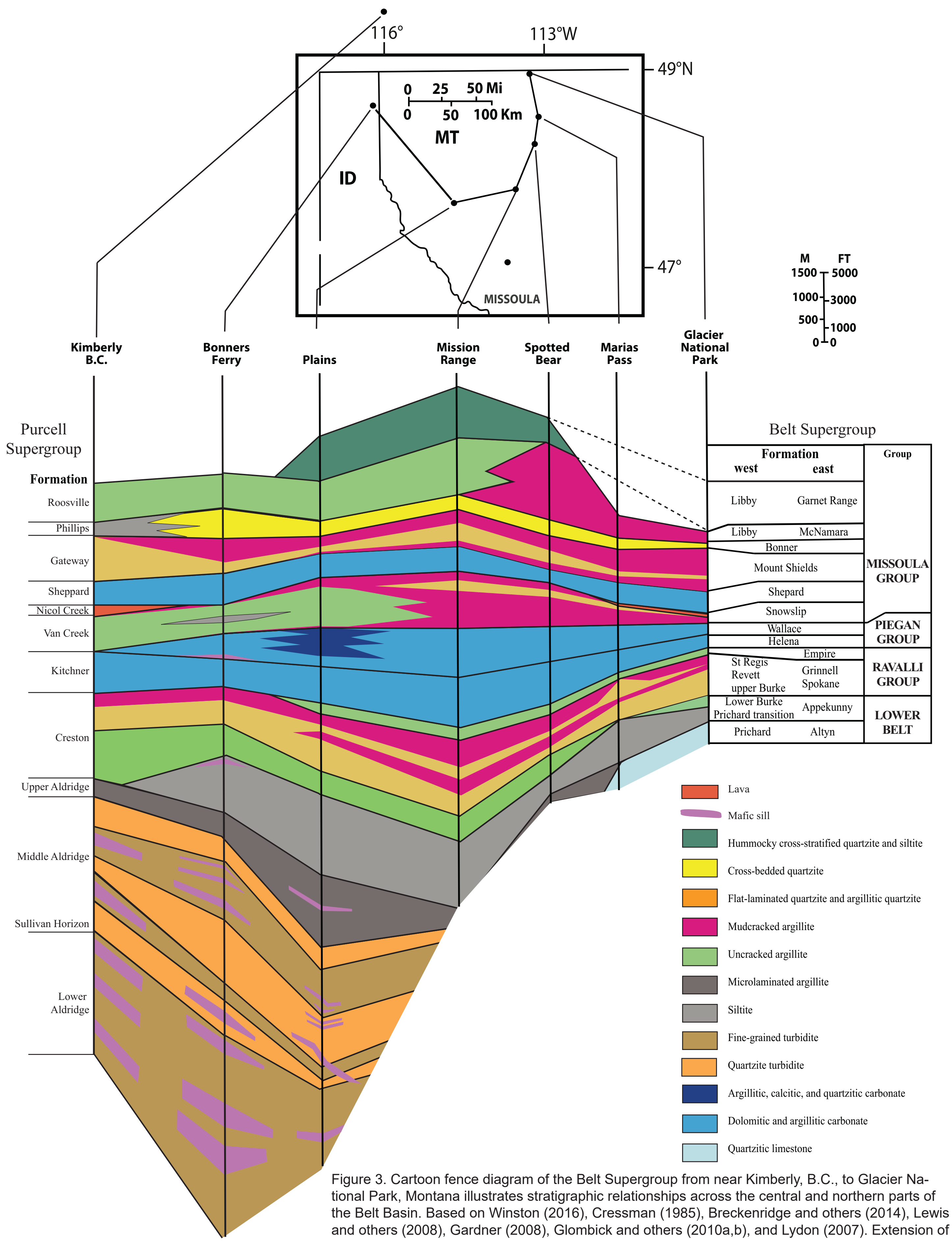


Figure 3. Cartoon fence diagram of the Belt Supergroup from near Kimberly, B.C., to Glacier National Park, Montana illustrates stratigraphic relationships across the central and northern parts of the Belt Basin. Based on Winston (2016), Cressman (1985), Breckenridge and others (2014), Lewis and others (2008), Gardner (2008), Glombick and others (2010a,b), and Lydon (2007). Extension of Prichard members C through G and Ravalli Group Formations into Canada by artistic license.

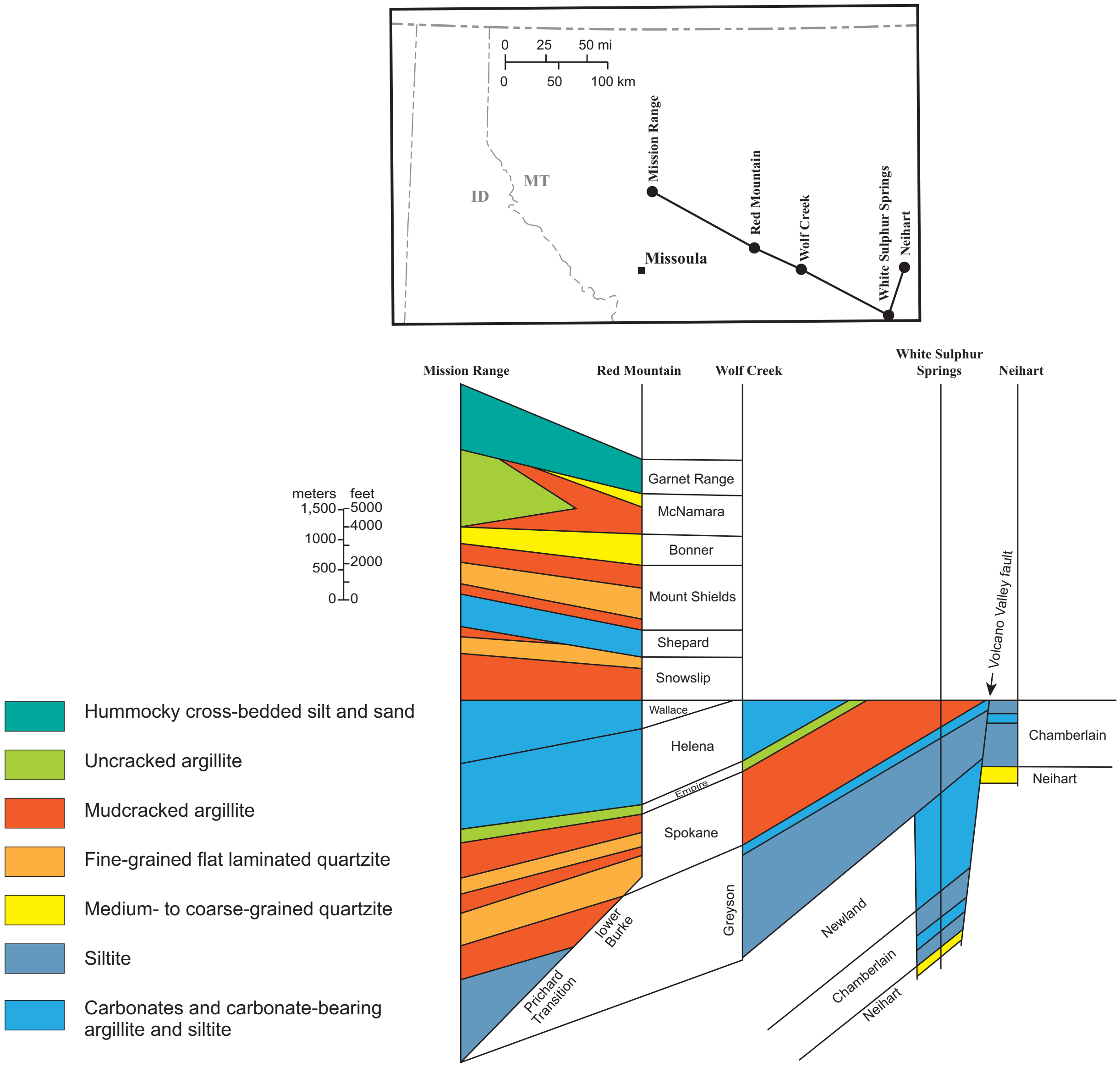


Figure 4. Cartoon fence diagram shows relationships between strata of the central Belt Basin and the Helena Embayment, an eastern arm of the Belt Sea. Based on figure 3 of this paper and Winston and Link, 1993.

part (Ross and others, 1992) are much too young to be Belt. Little work has been done on the Deer Trail Group since Winston and Link's (1993) summary, and we offer no updates to that review.

The "bottom" of the Belt is exposed in only a few areas (table 1). In these places, thin quartz arenite intervenes between the underlying Archean–Paleoproterozoic basement rocks and the overlying lower Belt strata (table 1). These underlying quartz arenites—the Neihart Formation in central Montana (Weed, 1899), the Gold Cup in northeastern Washington and northern Idaho (Doughty and Chamberlain, 2008; Buddington and others, 2016), the Fort Steele in Canada (Höy, 1982), and the Marble Creek in northern Idaho (Baldwin and others, 2016)—are thought to be remnants of a pre-Belt sandstone sheet that covered a continental-scale, northward-sloping pediment surface prior to the onset of Belt rifting (Sears, 2007a; Sears and Link, 2007; Ross and Villenueve, 2003). U-Pb detrital zircon plots (Ross and Villeneuve, 2003; Mueller and others, 2003, 2016; Doughty and Chamberlain, 2008; Buddington and others, 2016) from the orthoquartzite differ from Belt detrital zircon plots and appear to support this interpretation. Note that this and subsequent references to geographic directions are relative to our present coordinate system. At the time of Belt–Purcell deposition, the basin was in the southern hemisphere and north was to the present southwest (Elston and others, 2002).

Belt conglomerates rest depositionally on crystalline basement rocks along the southeastern Belt Basin margin in southwestern Montana (McMannis, 1963;

Pearson, 1996; McDonald and others, 2012; Tysdal, 2002). These Belt strata are of several stratigraphic levels and ages, and are interpreted to represent coarse, basin-margin facies deposited along the steep, possibly fault-controlled southeastern basin margin (McMannis, 1963; McDonald and others, 2012; McDonald and Lonn, 2013).

The western side of the Belt Basin is missing. Presumably it was rifted away and now resides on some other continent (summary in Hofmann and others, 2003), although it could simply have been subducted (Wallace, 1998). The various and conflicting interpretations of the paleogeographic and tectonic setting of the Belt Basin, and whether or not it was marine or lacustrine, are discussed in more detail later in this paper.

GROUPS OF THE BELT SUPERGROUP

Lower Belt

The basal strata of the Belt–Purcell Supergroup are traditionally designated as the lower Belt and assigned the stratigraphic rank of *group*. Although this nomenclature is informal, the tremendous thickness of the lower Belt, up to 12 km (Lydon, 2005; 7.5 mi) with another 5–7 km (3–4.3 mi) indicated by seismic imaging (Cook and van der Velden, 1995; Sears, 2013), warrants nothing less than *group* status. Along the eastern margin of the basin and well exposed in Glacier National Park, the lower Belt is represented by platform carbonates of the Altyn and Waterton Formations. However, the lower Belt quickly thickens westward across a north–northwest-striking syndepositional growth fault (Lydon and van Breeman, 2013), becoming the enormously thick (>6 km; >3.7 mi) Prichard Formation and its Canadian equivalent, the Aldridge Formation (fig. 3; table 1). Lydon and van Breeman (2013) postulated that the Prichard Formation is a rift-fill sequence composed of fine- and coarse-grained turbidite deposits interlayered with mafic sills, the oldest of which (1468 Ma) were emplaced into unconsolidated, wet sediments (Anderson and Davis, 1995; Buckley and Sears, 1998; Poage and others, 2000).

The Prichard Formation material was probably deposited within huge sandy alluvial aprons on the southwestern shore of the Belt Sea (Cressman, 1989; Lydon, 2005). The extent of this fan complex may have been nearly as large as the Mississippi River delta, but contained an even greater volume of sediment (Cressman, 1989). After reaching the Belt Sea, this material was carried northwestward, parallel to shore, by turbidity currents (Höy, 1993; Godard, 1998; Chandler, 2000; Lydon, 2005). The postulated southwestern source is supported by detrital zircon age peaks in the North American magmatic gap (NAMG) of 1625–1510 Ma that suggest a non-North American provenance (Ross and Villeneuve, 2003; Link and others, 2007; Lewis and others, 2010). However, lower Belt samples from the eastern side of the basin do not contain NAMG grains and are postulated to have come from the sea's eastern, Laurentian shores (Lydon and van Breeman, 2013).

Abundant pyrrhotite gives many Prichard outcrops a rusty red-brown color. The Prichard Formation also includes extraordinary intervals of millimeter-scale, alternating light and dark laminae that, incredibly, have been correlated lamina by lamina across the basin (summarized in Winston and Link, 1993). These silts must have settled out in deep water unaffected by turbidity currents. Indeed, most workers agree that the Prichard Formation was deposited in deep water in a restricted marine setting that was open to the ocean at least periodically (Winston and Link, 1993; Anderson and Davis, 1995; Lyons and others, 1998, 2000; Luepke and Lyons, 2001; Lydon, 2005). Also noteworthy are the very rapid accumulation rates shown in figure 2, with perhaps 1 km (0.6 mi) of sediment accumulating every million years. Future work will undoubtedly subdivide the Prichard into multiple formations, perhaps following present informal units (figs. 2, 3), but the current state of subunit definitions precludes widespread correlations. An exception is the "lined" unit (Prichard H, fig. 2) of even, parallel siltite and argillite that is the top of the Prichard Formation in Canada and most places to the south.

In the eastern Belt Basin, strata of the overlying Appekunny Formation constitute the uppermost part of the lower Belt, with correlations in progress to define the Appekunny westward across the basin and determine its relationship to the Burke Formation of the Ravalli Group (fig. 3). The Appekunny Formation and equivalent units are commonly purple and green, well-sorted, flat-laminated to very gently hummocky cross-stratified, very fine-grained sandstone and siltstone. The hummocks show that, unlike the underlying Prichard, it was deposited above wave base as the Belt Sea shallowed. As the waters became even shallower, an interval of very thinly laminated black pyritic mud accumulated at the top of the Appekunny (see discussion of microlaminated mud deposition in the Ravalli Group section).

Mineralogy suggests that the shallow waters of the Appekunny were fully oxygenated and supported aerobic eukaryotes (see discussion below), but pore waters in the underlying sediments were oxygen-poor and hydrogen sulfide-rich, promoting precipitation of pyrite (Slotznick, 2016).

Lower Belt in the Helena Embayment

The Helena Embayment, a narrow eastern arm of the Belt Basin (fig. 1), contains lower Belt strata that differ from those of the central basin and have a different nomenclature (fig. 4, table 1). They are correlated with the lower Belt on the basis of stratigraphic position beneath the Ravalli Group (fig. 4), but no absolute age data are available.

In the Helena Embayment, the basal Chamberlain Formation overlies the pre-Belt(?) Neihart Quartzite discussed earlier. The Chamberlain Formation is black shale and hummocky silt and arenite (Winston and Link, 1993), with microfossils indicating a very shallow marine environment (Adam and others, 2016). It grades upward to the Newland Formation, which is as thick as 3,000 m (1.9 mi) and composed of parallel-laminated shale interbedded with debris flows, turbidites, and carbonates that are more common toward the top (Slotznik and others, 2015). The overlying Greyson Formation is wavy laminated siltstone and dark shale with interbedded turbidites (Whipple and Morrison, 1993) and a microfossil assemblage consistent with distal shelf marine environments (Adam and others, 2014, 2016). Like the lower Belt in the central basin, Greyson Formation strata are mostly deeper water sediments. Not shown in figure 4 is the LaHood Formation, a wedge of coarse conglomerate deposited by debris flows and turbidites along the fault-bounded southern edge of the Helena Embayment (McMannis, 1963). Its limited outcrop extent makes correlations with other lower Belt units uncertain, although it has been described as interfingering with the other Helena Embayment units (O'Neill, 1995, 1998; Ross and Villeneuve, 2003; McDonald and others, 2012; McDonald and Lonn, 2013). It is thought to be a basin margin facies developed adjacent to a major east–west syndepositional normal fault (McMannis, 1963). However, detrital zircons show that the LaHood Formation had a different provenance than the other eastern lower Belt units discussed above (Mueller and others, 2016; Guerrerro and others, 2016), so either there were numerous fault-bounded, isolated subbasins with local sources or the LaHood is older than the Belt.

Ravalli Group

Stratigraphically above the lower Belt, purple and green mudcracked argillite marks the base of the Burke, Spokane, and Grinnell Formations at the bottom of the Ravalli Group (figs. 3, 4). These mudcracked strata reflect deposition in increasingly shallow waters and mark the change from perennial subaqueous deposition of the lower Belt to episodic subaerial exposure in the overlying Ravalli Group. The alternating centimeter-scale layers of siltstone overlain by commonly desiccation-cracked mudrock formed as each depositional event first dropped silt from its bed load and then mud when the clay settled out of suspension. When the floods ebbed and the sea shrank, the mud caps dried out and cracked (Winston and Link, 1993). Each centimeter-scale siltite to argillite layer has been termed a couplet by Winston (1986c; thicker decimeter-scale sand/silt/clay fining-upward layers are termed couples, and thinner millimeter-scale layers are microcouplets). Each couplet or couple is interpreted to represent a single depositional event.

In the western basin, the silt–clay couplets at the bottom of the Ravalli Group grade up to thick intervals of fine-grained sand characteristic of the Revett Formation (fig. 3). In the eastern basin, the entire Ravalli Group section is composed of mudcracked red siltite and argillite, the Grinnell and Spokane Formations (figs. 3, 4) that laterally grade westward into the sandy Revett Formation. Winston (2016) analyzed the sedimentology of the Revett by identifying sedimentary structures and interpreting the flow processes that created them. He described typical event beds in the Revett Formation as:

> "characterized by tabular, flat-laminated, arenite beds from decimeters up to a meter thick. Trough and planar crossbeds lie below the flat-laminated layers in many outcrops and form the bases of upward-thinning and -fining successions. The flat-laminated, fine-grained arenite layers commonly pass upward to three dimensional flow ripple crossbeds that in turn grade upward to thin, mudcracked argillite layers. In this succession, flow that deposited the large basal crossbeds began in the upper part of the lower flow regime. The overlying flat-laminated arenite layers show that flow shifted to the upper regime, probably as flow shallowed to less than a meter deep. The overlying three-dimensional flow ripples record

> shift to the lower part of the lower flow regime, and the desiccation-cracked mud indicates that the flow stopped and the surfaces dried. Clearly these vertical successions imply decelerating flow of repeated floods.” (Winston, 2013, p. 73).

Laterally, tongues of Revett quartzite pinch out eastward over a distance of more than 200 km (124 mi) into the thin siltite and mudcracked argillite beds of the Grinnell and Spokane Formations (figs. 3, 4) near the basin center. The trough-crossbeds at the base of some Revett beds resemble channels in cross-sectional view, but Winston (2016) instead interpreted them as scour pits filled by dunes (fig. 5) because they are not associated with the accretionary crossbeds typically found in channel deposits. Therefore, Winston (2013, 2016) interpreted that Revett was deposited by unconfined, sandy sheetfloods that flowed eastward across broad alluvial megafans. These megafans were of titanic dimensions, unlike any on modern Earth. The sheetfloods flowed at grade rather than cutting channels into the underlying deposits. The sand tongues pinched out basinward as the water seeped into the underlying sand and dropped its load, and as the sheetfloods began to pond near the basin center (Winston, 2016). As the floods crossed dry mudflats, they deposited graded sand to mud couples and couplets. Farther on, they ponded, depositing silt beds sharply overlain by mud that settled from suspension. Desiccation cracks formed in the mud caps following the floods, and when the sea receded. The sheetflood model can also be applied to the sandy units of the Missoula Group and Lemhi strata in the upper part of the Belt (Winston, 2013; Winston and Sears, 2013).

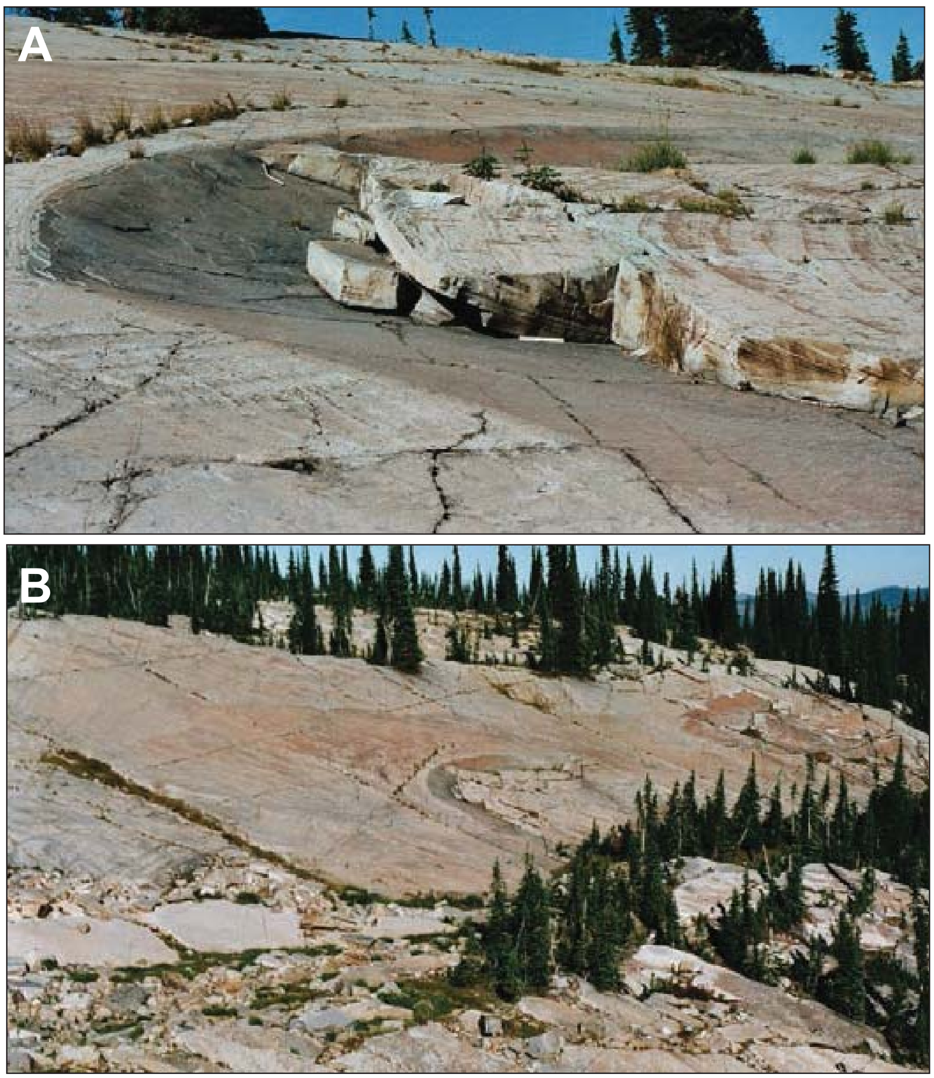


Figure 5.Trough crossbeds filling scour pits. (A) Large trough crossbed, 16 m across, was interpreted by Winston (2016) as filling a scour pit eroded into the underlying flat-laminated sand beds. Crossbeds were deposited by a three-dimensional flow dune in water that was deep enough to flow in the upper part of the lower-flow regime. Flow was from the upper left. The crest of the dune was flattened, probably as a flood waned, shallowed, and shifted to upper-regime plane bed flow. White bar in center is 30 cm. From Winston, 2016. (B) Scour pit of A flanked on the right and left by equally large trough crossbeds (red arrows), showing that flood flow formed a sheet of large three-dimensional flow dunes. From Winston, 2016.

Only two U-Pb detrital zircon samples are available for the Ravalli Group (Ross and Villeneuve, 2003; Balgord and others, 2008), but because they yielded significant grains from the North American magmatic gap (1625–1510 Ma), they appear to confirm the postulated western, non-North American provenance (Winston and Link, 1993; Winston, 2016).

The finer-grained eastern Ravalli Group facies (Grinnell and Spokane Formations) contain thin beds of quartz-rich, coarse- to medium-grained, well-rounded sand. These beds of clean, coarse sand pinch out westward, and are much different than the fine-grained, feldspathic Revett quartzites that pinch out eastward. They are postulated to have originated on Laurentia to the east rather than from a western source like the Revett Formation sands (Harrison and others, 1997; Winston and Sears, 2013), providing evidence that the Belt Basin was intracratonic and possibly a giant, landlocked, playa lake. Pratt’s (2001, 2017) contrasting interpretation is that the mudcracked

Grinnell Formation (eastern Ravalli Group facies) was deposited in deep marine waters. Because some mudcracks contain mudchips that appear to have been injected upward, he postulated that they were not formed by desiccation but instead are syneresis cracks formed through dewatering or generated by seismic activity. The coarse-grained beds of clean sand represent eastern shore beach sands that were reworked by tsunamis. The question of desiccation cracks versus syneresis cracks is examined later in this paper.

No other recent sedimentological work on the Ravalli Group is available. However, prior to 1993, the marine versus lacustrine debate for the Ravalli Group and the rest of the Belt strata was contentious (see Winston and Link, 1993, and references therein). Also, some lithologies in the Lemhi Subbasin that are similar to the Revett Formation are interpreted as marine tidal flat–delta floodplain deposits (Tysdal, 2000). Obviously, the question of marine versus lacustrine deposition remains unsettled.

The top of the Ravalli Group is characterized by mudcracked siltite-to-argillite couplets with interbedded centimeter- to decimeter-scale tabular sandstone beds. These grade upward to green, uncracked couplets across the entire basin, apparently marking the expansion of the Belt Sea.

The Empire Formation has been mapped at the top of the Ravalli Group in the eastern part of the basin and in Canada (Binda and Koopman, 1998), but is not recognized in the west. It appears to be a transitional unit into the subaqueous depositional environment interpreted for the overlying Piegan Group.

Piegan Group

The Piegan Group, formerly called "middle Belt carbonate," contains abundant dolomite and calcite, making it more recognizable than most Belt units. The Piegan Group is also characterized by hummocky cross-stratified sandstone and siltstone beds and abundant oscillation ripples that suggest deposition in deeper water than the underlying Ravalli Group, but water still shallow enough that the bottom was within reach of storm waves. The Helena and overlying Wallace Formations (table 1) that constitute the Piegan Group (figs. 3, 4) in the U.S. were thoroughly studied by Winston (2003, 2007). He reported that both are characterized by cycles in which hummocky lenses of very fine sand or silt are gradationally overlain by mud, and these graded couples and couplets become thinner and finer grained both upward within the cycles, and laterally near the eastern margin of the basin. In the Helena Formation, dolomite is abundant in the upper parts of the cycles, and full cycles are capped by very thinly laminated, millimeter-scale, silt-to-clay microcouplets. The dolomite was precipitated from supersaturated water, and, contrary to traditional sedimentological models, the microcouplets at the top of cycles and in the eastern basin appear to be very shallow water deposits. Some are mudcracked, and Winston (2003, 2007) postulated that they formed when storms drove sediment-laden water across shallow mudflats. Johnson (2013) agreed that the microcouplets were deposited on shallow mudflats, but favors deposition by tides. The Wallace Formation has similar, but thicker and less obvious, cycles. The thicker (decimeter-scale) hummocky cross-stratified beds suggest the water was deeper during Wallace deposition, but the unit displays similar fining and thinning upward and eastward. Calcite is more common than dolomite, but some parts are devoid of carbonate. Near the eastern margin of the basin, oolitic grainstones are present.

Winston (2003, 2007) attributed the upward and lateral thinning and fining cycles in both formations to shallowing and shrinking of the Belt Lake during arid times. He interpreted the dolomite and calcite as chemical deposits, possibly biologically induced, that precipitated when the evaporating playa lake became supersaturated. The carbonate-poor parts of the Wallace Formation may record deposition in a balanced-fill lake, which had an outlet at higher levels. The eastward thinning and fining was interpreted to be the result of shallower water near the eastern, low-relief margin of the basin.

Pratt (1998b, 2001), who has carefully studied equivalent Piegan Group strata in Canada (table 1), has a contrasting interpretation of the depositional environment there (fig. 6). He contended that the Piegan Group was deposited in a vast epeiric sea that extended well east of the current Piegan Group exposures. The sea was deep enough to limit storm-induced turbulence of the sediments except for occasional high-energy events generated by tsunamis. Tysdal (2003) also proposed deep-water origins for similar, Wallace-like lithologies in the Lemhi Subbasin (Apple Creek Formation, banded siltite member), interpreting them as turbidites. However, although there were other proponents of a marine environment for the Piegan Group, most thought it was shallow. Based on isoto-

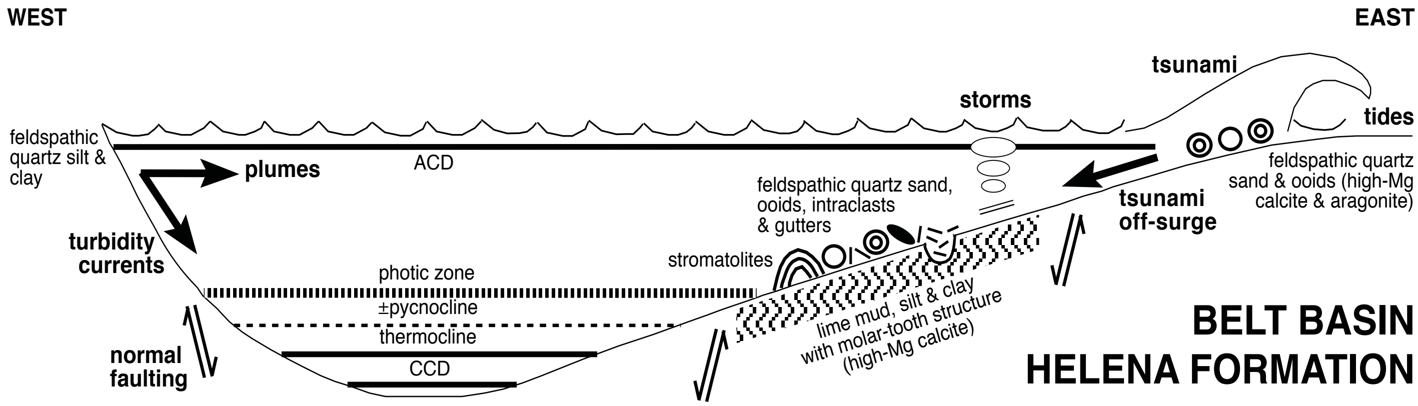


Figure 6. Pratt's (2001) east–west profile across the Belt Basin during deposition of the Helena Formation, Piegan Group, shows deep marine waters in contrast to Winston's interpretation of a shallow playa lake. Not to scale; the western side of the basin floor was of imperceptible relief and slope, and largely within the photic zone and above the carbonate compensation depth and thermocline. The dominant sources for terrigenous sediment were tectonically active areas to the west and southwest with material transported via turbidity currents and plumes. High-Mg calcite lime mud precipitated in situ, probably in the water column. A pycnocline developed occasionally and at times was elevated into fairly shallow water. The basin was likely microtidal. Erosion has removed the eastern shore of the basin, but it had shoals of high-Mg calcite and locally aragonitic ooids, and probably tidal flats that at times were evaporitic. Normal faulting on the western side generated tsunamis that swept onto the shallow eastern side of the basin. Tsunami off-surge formed west-directed currents that transported ooids into deeper water and created gutters, but storm-induced reworking was relatively weak. Frequent synsedimentary faulting during subsidence caused dewatering and lime mud injection, resulting in ubiquitous molar-tooth structure. From Pratt, 2001, with permission.

pic work, Frank and others (1997) postulated that the Helena Formation was a shallow marine basin that was increasingly isolated from the open ocean. Johnson (2013) proposed that the Piegan Group represents a shallow tidal environment dominated by fluid mud. As for the Ravalli Group, the marine versus lacustrine debate continues.

The only two U-Pb detrital zircon samples from the Wallace Formation (Ross and Villeneuve, 2003; Lewis and others, 2007) contained non-North American grain age populations, confirming a generally accepted western source for some of the Piegan Group material (Winston and Link, 1993; Wallace, 1998; Höy, 1993; Pratt, 2001; Winston, 2003, 2007). However, a sample from the equivalent Siyeh Formation in Glacier National Park (Lydon and van Breeman, 2013) had an Archean peak instead, suggesting that the eastern Piegan Group had an eastern provenance on Laurentia. Therefore, like the strata of the lower Belt and Ravalli Group, Piegan Group sediments likely had sources to both the west and the east.

There are several geologic features common in the Piegan Group that have attracted in-depth study and bear on interpretations of the depositional environment and tectonic setting. Crinkle cracks, molar-tooth structures, and the Wallace breccia are discussed below.

Crinkle Cracks

The term crinkle cracks was coined by Winston (2013) for discontinuous, non-intersecting, commonly curved and spindle-shaped, sand- or silt-filled mud cracks that are contorted in cross-sectional view (fig. 7). They are sometimes preferentially oriented into subparallel arrays. They are distinct from polygonal, straight-sided sand-filled desiccation cracks. Crinkle cracks are interpreted as a type of syneresis crack, which forms subaqueously and is generally ascribed to dewatering. Crinkle cracks commonly occur in the Piegan Group, but can be seen in other Belt strata, and are also widespread in Proterozoic and Phanerozoic marine and lacustrine rocks (e.g., Winston and Smith, 2016 and references therein). Based on fieldwork on a Louisiana coastal mud bank where crinkle cracks appear to be actively forming, Winston and Smith (2016) proposed that when waves strike viscous, fluid mud beds, they transfer their energy to solitary compressional waves that open cracks in the mud on their trailing edges that then fill with transported sand. When the fluid mud is later buried and vertically compressed and the water is driven out, the cracks become contorted, or crinkled, in cross section.

In contrast, Pratt (1998a, 2001, 2017a,b) and Pratt and Ponce (2019) described cracks filled with breccia-like intraclasts (fig. 7) that they attributed to dewatering during seismic activity, and attributed most cracks, including crinkle and some "desiccation" cracks, to seismicity. Winston and Smith (2016) rejected the seismic hypothesis, pointing out that cracks are not characteristic of seismites, and that the Piegan Group lacks any commonly accepted characteristics of seismites.

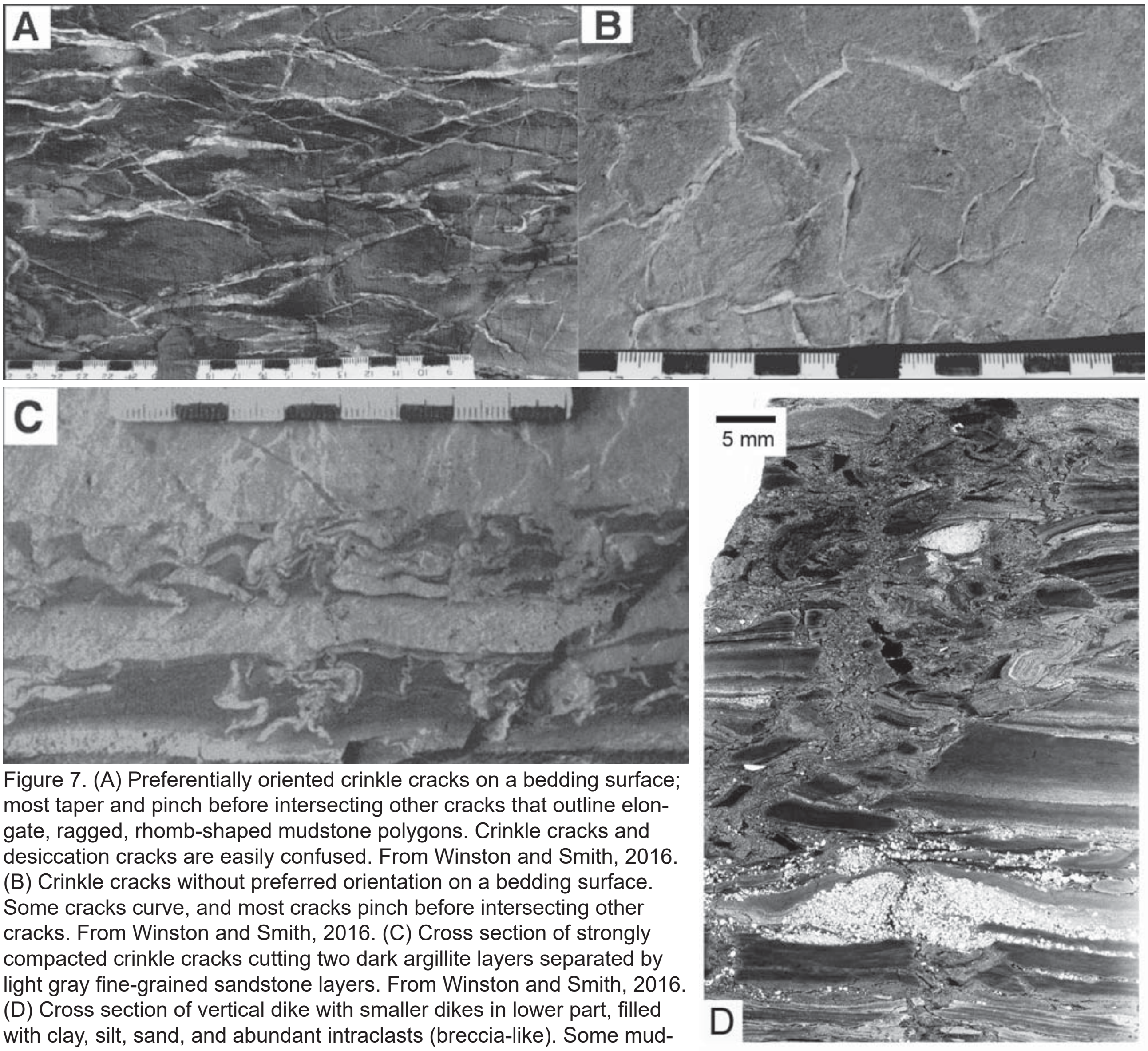


Figure 7. (A) Preferentially oriented crinkle cracks on a bedding surface; most taper and pinch before intersecting other cracks that outline elongate, ragged, rhomb-shaped mudstone polygons. Crinkle cracks and desiccation cracks are easily confused. From Winston and Smith, 2016. (B) Crinkle cracks without preferred orientation on a bedding surface. Some cracks curve, and most cracks pinch before intersecting other cracks. From Winston and Smith, 2016. (C) Cross section of strongly compacted crinkle cracks cutting two dark argillite layers separated by light gray fine-grained sandstone layers. From Winston and Smith, 2016. (D) Cross section of vertical dike with smaller dikes in lower part, filled with clay, silt, sand, and abundant intraclasts (breccia-like). Some mudstone fragments are folded (e.g., center right). These resemble neither crinkle nor desiccation cracks, and were interpreted by Pratt (1998a, 2001, 2017a,b; Pratt and Ponce, 2019) as subaqueous muds rocked by earthquakes. From Pratt and Ponce, 2019.

Harazim and others (2013) attributed crinkle cracks to volume reduction of microbial stabilized mud via fluid removal, but clearly the cracks formed *before* fluid removal and compaction (Winston and Smith, 2016).

Molar-Tooth Structures

Molar-tooth structures (MTS) are vertical or horizontal calcite ribbons and blobs, often forming interconnected networks (fig. 8). MTS are widespread in the Piegan Group and other Precambrian sediments but are unknown in the Phanerozoic section. Smith (2016) reviewed previous work and speculated on their origin, which is fortunate because he reported more than 300 publications on MTS. The vertical calcite ribbons commonly are highly contorted, similar to crinkle cracks in cross section, but the horizontal ribbons are not. This is consistent with calcite precipitation occurring while the mud was still fluid and before it was compacted. The tops of some ribbons are scoured, showing that they developed at depths of less than 1 m (3.3 ft; Furniss and others, 1998; Pratt, 1998b, 2001). Calcite fragments in the bases of Helena cycles are interpreted to have resulted from this scouring. MTS are most common in clayey calcitic or dolomitic microspar and silty dolomite in the upper halves of the Piegan Group cycles (Furniss and others, 1998). The calcite must have filled voids developed in the uncompacted mud, and the origin of those voids has been the subject of much study. Perhaps because the vertical ribbons share the same tightly folded shape as syneresis cracks, syneresis (intrastratal) cracks with

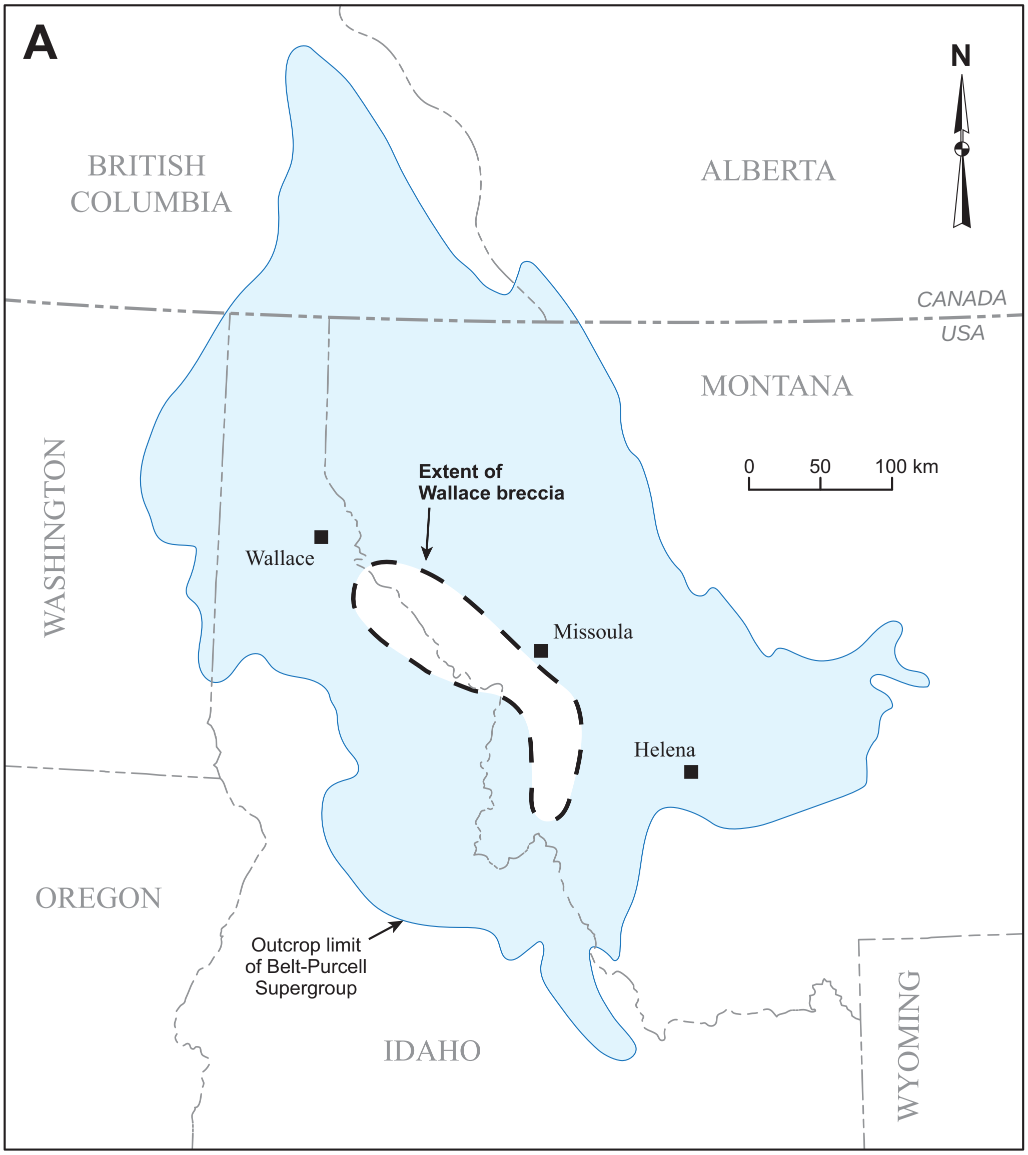
Figure 8. Cross-sectional view of molar-tooth structures, calcite-filled ribbons that are common in carbonate-mud beds of the Piegan Group. While they are common in Precambrian sediments, they are unknown in the Phanerozoic. Note pencil for scale.

many proposed origins (e.g., Smith, 2016 and references therein) have been postulated as the precursors. However, Smith (2016) pointed out that intrastratal cracks in fluid, uncompacted mud are likely to be immediately filled with material from the surrounding or overlying sediment, not with calcite precipitate. In addition, intrastratal cracks explain neither the blob-shaped nor the horizontal MTS.

The mostly commonly accepted hypothesis is the gas expansion crack model, which proposed that gas produced from the decay of microbial mats created the voids (Furniss and others, 1998; Lyons and others, 2003; Bishop and others, 2006; Pollock and others, 2003, 2006). This process explains the common geometry of blobs linked by networks of ribbons and is supported by experimental data. Furniss and others (1998) generated CO_2 using yeast and sugar in water-saturated kaolinite. The larger bubbles rose upward through the mud, leaving water-filled pathways behind that then served as outlets for subsequent bubbles. The rising bubbles pumped water from the clay, stiffening it. When the surface was sealed, the confined gas formed jagged, sinuous, downward, upward, and horizontal cracks that cut through the earlier formed bubbles. The result was a series of pathways that closely mimicked common MTS geometry. The same processes probably occurred in the Precambrian sediments, with the surface seal formed by microbial mats. Subsequent precipitation of the microspar calcite was probably complex, and may have involved recrystallization of initially precipitated amorphous calcium carbonate or vaterite (Gellatly and Winston, 1998; Gellatly and Lyons, 2005).

Wallace Breccia

Enigmatic breccia in the Wallace Formation may have implications for the depositional environments or tectonic setting of the basin. The breccia crops out in an arcuate zone 150 km (93 mi) long in western Montana and northern Idaho (fig. 9) at roughly the same stratigraphic level in the upper part of the Wallace Formation, usually within intervals of thin hummocky white silt beds that grade upward to tan, carbon-



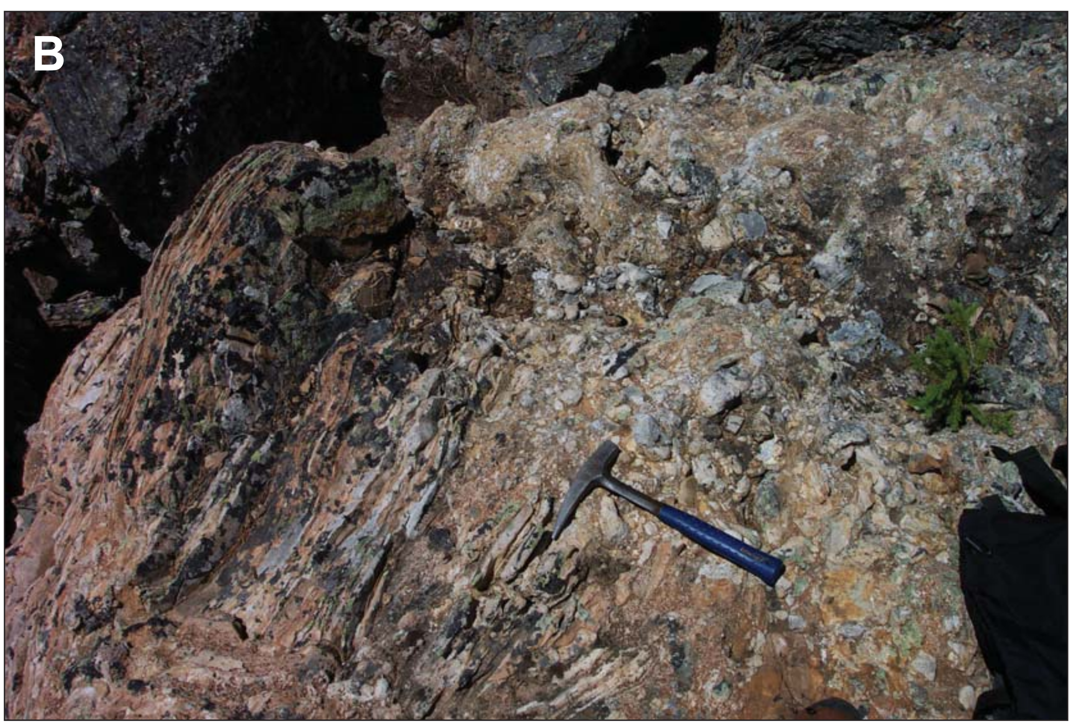


Figure 9. Wallace breccia is stratabound and sometimes stratiform. (A) Wallace breccia occurs in an arcuate zone near the Idaho–Montana border. Modified from Overocker, 2006. (B) Wallace breccia clasts are angular and highly variable in size; clasts can be as large as 10 m across.

ate-bearing mud. The breccia is both concordant and discordant to enclosing beds, is locally associated with tight folds, and contains equant, elongate, rounded, angular, and distorted clasts ranging in size to more than 10 m (33 ft). Possible origins include: (1) syndepositional downslope slumping (Godlewski, 1981; Thiessen and Wallace, 1996; Overocker, 2006); (2) syndepositional evaporite dissolution and collapse (Campbell, 1960); (3) post-depositional tectonic faulting (Collins and Winston, 1999); and (4) hydraulic fracturing associated with pluton emplacement (Overocker, 2006). Because the breccias are stratabound, it seems unlikely that they have several different origins as proposed by Overocker (2006). Syndepositional downslope slumping is rejected because some clasts are angular, and therefore post-lithification, and there is no evidence for significant slopes in the Wallace: Wallace lithologies are uniformly shallow water facies deposited on a flat sea floor (Winston, 2003, 2007). Collins and Winston (1999) thought that the breccias resulted from post-depositional faults, possibly of Proterozoic age, that crossed strata at low angles. Burmester and others (1998) noted that the west–northwest-elongated area of breccia (fig. 9A) parallels regional faults, suggesting a genetic relationship. Lonn (2014) speculated that, because their occurrence roughly outlines the boundaries of the Lemhi Subbasin (fig. 1), they may have developed during the rapid subsidence of the Lemhi Subbasin, which accumulated more than 14 km (8.7 mi) of sediment over roughly 50 million years (see Lemhi Subbasin section below). But if the breccia is fault-related, why does it not extend into overlying Missoula Group or underlying Wallace and Helena strata? The breccia commonly contains abundant scapolite crystals (Campbell, 1960; Lonn and McFaddan, 1999), a sodium- and chlorine-rich metamorphic mineral that can originate from evaporites (e.g., Kříbek and others, 1997), so perhaps the evaporite dissolution and collapse origin (Campbell, 1960) should be reexamined.

Missoula Group

Following the mostly subaqueous deposition of the Piegan Group, the Belt Sea shallowed and experienced episodic regressions and transgressions, resulting in the alternating fluvial and shallow water deposits that dominate the Missoula Group stratigraphic column (fig. 3). The Missoula Group has been better-studied than the underlying strata, perhaps because it is more colorful and lithologically diverse. The diverse lithologies reflect their locations on the alluvial megafans and within the shallow, fluctuating Belt Sea. Seaward locations are characterized by mudcracked siltite–argillite couplets that record subaerial exposure at the end of each depositional event, or uncracked, sometimes dolomitic, couplets that record perennial shallow, subaqueous deposition. Higher on the megafans are fluvial deposits of thicker-bedded, fine- to coarse-grained sand. Therefore, the Missoula Group records depositional environments similar to those of the Ravalli Group, and proposed depositional models are similar (for a more detailed discussion, see the Ravalli Group section above). Winston's (1986a,b,c; Winston and Link, 1993) playa lake model is less contentious for the Missoula Group than for underlying strata, although that may simply reflect the attrition of marine proponents working on the Missoula Group.

All formations in the Missoula Group become increasingly sandy and thicker-bedded southward (fig. 10; Winston, 1986a; Lydon, 2007), suggesting a southern provenance (fig. 11). In addition, Missoula Group U-Pb detrital zircon samples lack the westerly-derived, non-North American grains (Ross and others, 1992; Link and Fanning, 2003; Ross and Villeneuve, 2003; Gardner, 2008; Stewart and others 2010; Lydon and van Breeman, 2013; Hendrix and others, 2016) found in older Belt strata. Missoula Group detrital zircon data, with peaks from 1700 to 1800 Ma, do not match local basement rocks exposed to the south and east, and therefore a distant source has been proposed, such as the Mazatzal/Yavapai orogenic belt (late Paleoproterozoic) of the present day southwestern U.S. (Link and others, 2007, 2013, 2016; Sears, 2007b; Stewart and others, 2010). Sears (2007b) postulated that the Mazatzal/Yavapai grains were stored as a veneer on and transported across a continental-scale, northward-sloping pediment surface (fig. 12). However, Mueller and others (2016) suggested that these grains may have had a more proximal source in plutons associated with the 1.79–1.72 Ga Big Sky orogeny, which may have been well exposed in Belt time but are restricted in outcrop areas today.

These detrital zircon data together with the southward coarsening grain size suggest that the beginning of Missoula Group deposition represents a major change in the configuration of the Belt Basin. It appears that Missoula Group sediments were deposited by sheet floods flowing northward and northwestward across enormous megafans to pond against the tecton-

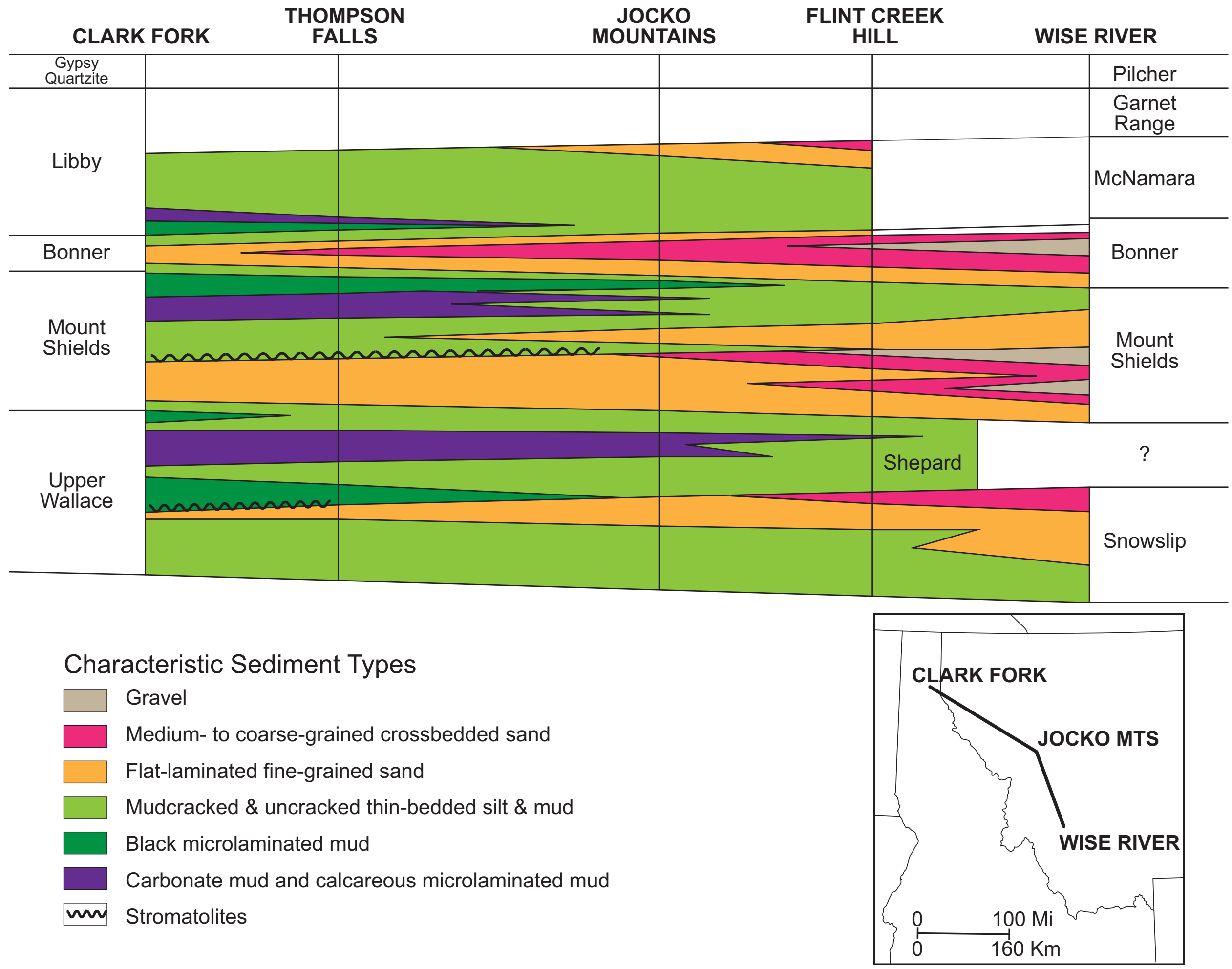


Figure 10. North–south fence diagram of Missoula Group lithologies shows how the sand wedges pinch out basinward and illustrates facies changes that occurred along the alluvial aprons that originated to the south (right). Note that the microlaminated mud is interpreted to represent deposition on very shallow mudflats. Modified from Winston, 1986a.

ically quiescent Laurentian land mass to the north and northeast (Winston, 1986a, 2013; Winston and Link, 1993; fig. 11). The contrasting marine-delta interpretation (Wallace, 1998) proposed that the sediments were deposited in a northward-prograding river delta system entering the open ocean.

A Few Formations in the Missoula Group Need Further Discussion:

In the Coeur d'Alene mining district, Snowslip and Shepard Formation correlatives were originally assigned to the upper Wallace Formation based on the presence of dolomite and very thinly laminated black argillite (Ransome and Calkins, 1908). Subsequent workers there (Lewis and others, 2008; Browne, 2006, 2017; Breckenridge and others, 2014) now accept Winston's Snowslip–Shepard correlations (Lemoine and Winston, 1986; Winston and Wheeler, 2006; table 1).

Northwest of Missoula, in an unusually dolomitic Snowslip Formation section, Lonn (2011) observed that dolomite content is high in all sediment types, including the sheetflood sands, and therefore attributed it to chemical precipitation from pore waters during diagenesis. The eastern Snowslip Formation contains lenses of white, coarse-grained quartzite (Lonn, 2015) similar to those of the eastern Ravalli Group that are interpreted to have an eastern Laurentian provenance.

The Shepard Formation, characterized by thin-bedded dolomitic green siltite and argillite, varies little in lithology or thickness across hundreds of kilometers, and is an important marker unit for Belt mappers (fig. 10). It appears to record mainly perennial subaqueous deposition in a shallow body of water.

The Bonner Formation contains the coarsest grains of any Belt unit and is conglomeratic in the south (fig. 10). In its type locality near Missoula, it is characterized by tabular sheets of trough crossbeds "forming a gallery of smiley-face lenses" (Winston and Sears, 2013, p. 264), capped by finer, thinner, flat-laminated beds. While it is tempting to interpret the trough crossbeds as channels within braided stream depos-

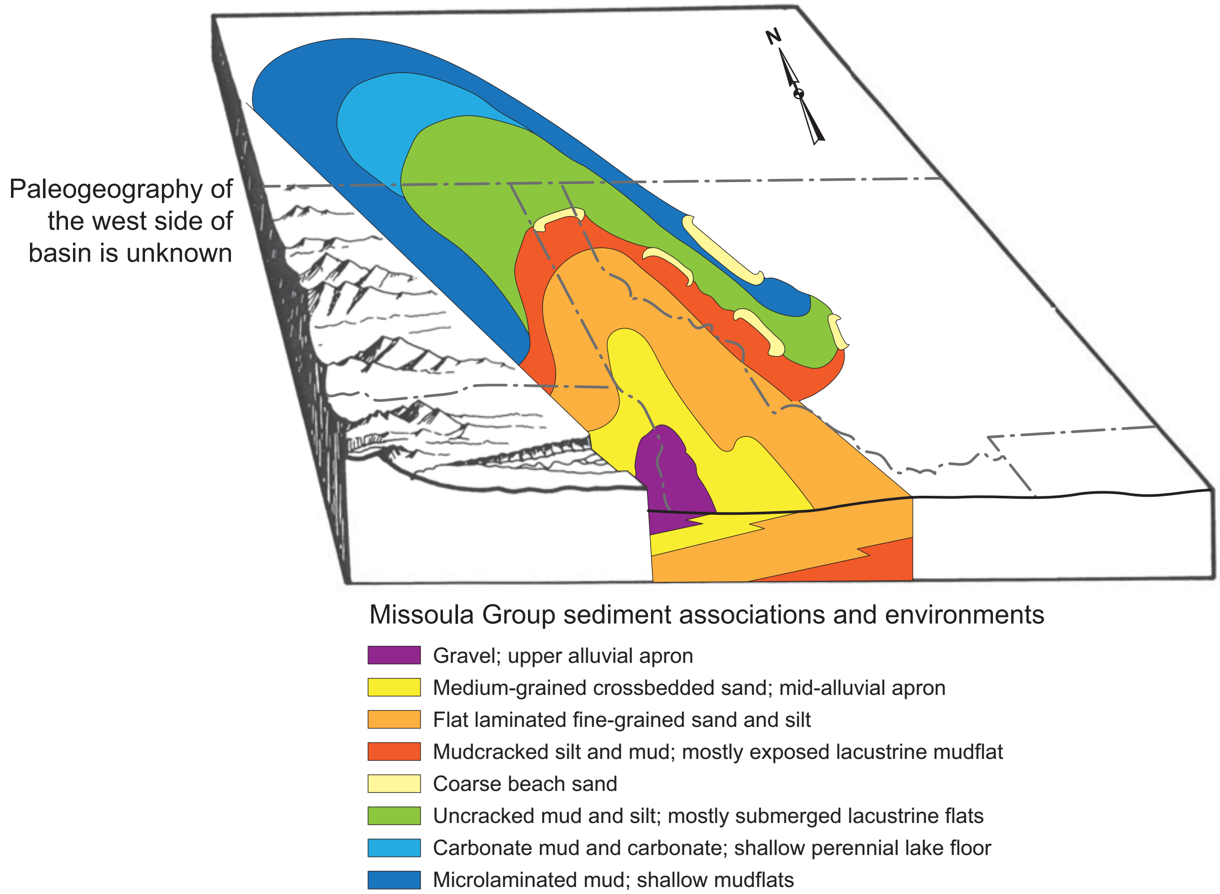


Figure 11. Block diagram shows facies associations during deposition of the Missoula Group. Sand and gravel facies alternately prograded and retreated, bordered by exposed mudflats of silt and desiccation-cracked mud that passes seaward to uncracked silt and mud and carbonate-bearing mud. Note that microlaminated mud was deposited on shallow mudflats near the shores of the Belt Sea. Modified from Winston, 1986a.

its, Winston cites the lack of channel edges and the continuity of the finer-grained tops to argue that they formed as scour pits (fig. 5) during northwest-flowing sheet floods (Winston and Sears, 2013). The thinner bedded, finer-grained, flat-laminated caps were deposited as flow waned at the end of each flood event. Winston and Sears (2013) caution geologists not to interpret Proterozoic deposits using Phanerozoic models because the Proterozoic landscapes probably looked nothing like Phanerozoic ones.

The Garnet Range and equivalent upper Libby Formations are characterized by hummocky crossbeds deposited in a Belt Sea that rapidly deepened in latest Belt time (Kidder, 1998). These sediments, like those of the Piegan Group, accumulated in a perennial water body where the depth was within reach of storm waves.

Strata of the Lemhi Subbasin

Mesoproterozoic sedimentary rocks of east-central Idaho, the Lemhi Group and related strata (table 1), consist of a monotonous succession of mainly feldspathic quartzite as much as 14 km (9 mi) thick (Burmester and others, 2013, 2016). A plutonic–metamorphic complex separates these strata from the Belt strata, and so one interpretation was that they were deposited in separate basins that were later tectonically juxtaposed by major structures (Ruppel, 1993; Ruppel and others, 1993; Evans and Green, 2003; Ruppel and O’Neill, 2003; O’Neill and others, 2007). Concurrently, Winston and others (1999) proposed direct correlations between Belt and Lemhi strata, suggesting only one basin. More recently, a >10-year ongoing collaboration between the Idaho Geological Survey (IGS) and the Montana Bureau of Mines and Geology (MBMG) built on copious previous work (Ruppel and others, 1993; Tysdal, 1996a,b, 2000a, 2002, 2003; Tysdal and Moye, 1996; Tysdal and others, 2005; Evans and Green, 2003, and references therein), resulting in new 1:24,000-scale mapping that was focused on revision of the Lemhi stratigraphic column. This work suggested adding 4,500 m (2.8 mi) of strata to the top of the previously recognized section (Burm-

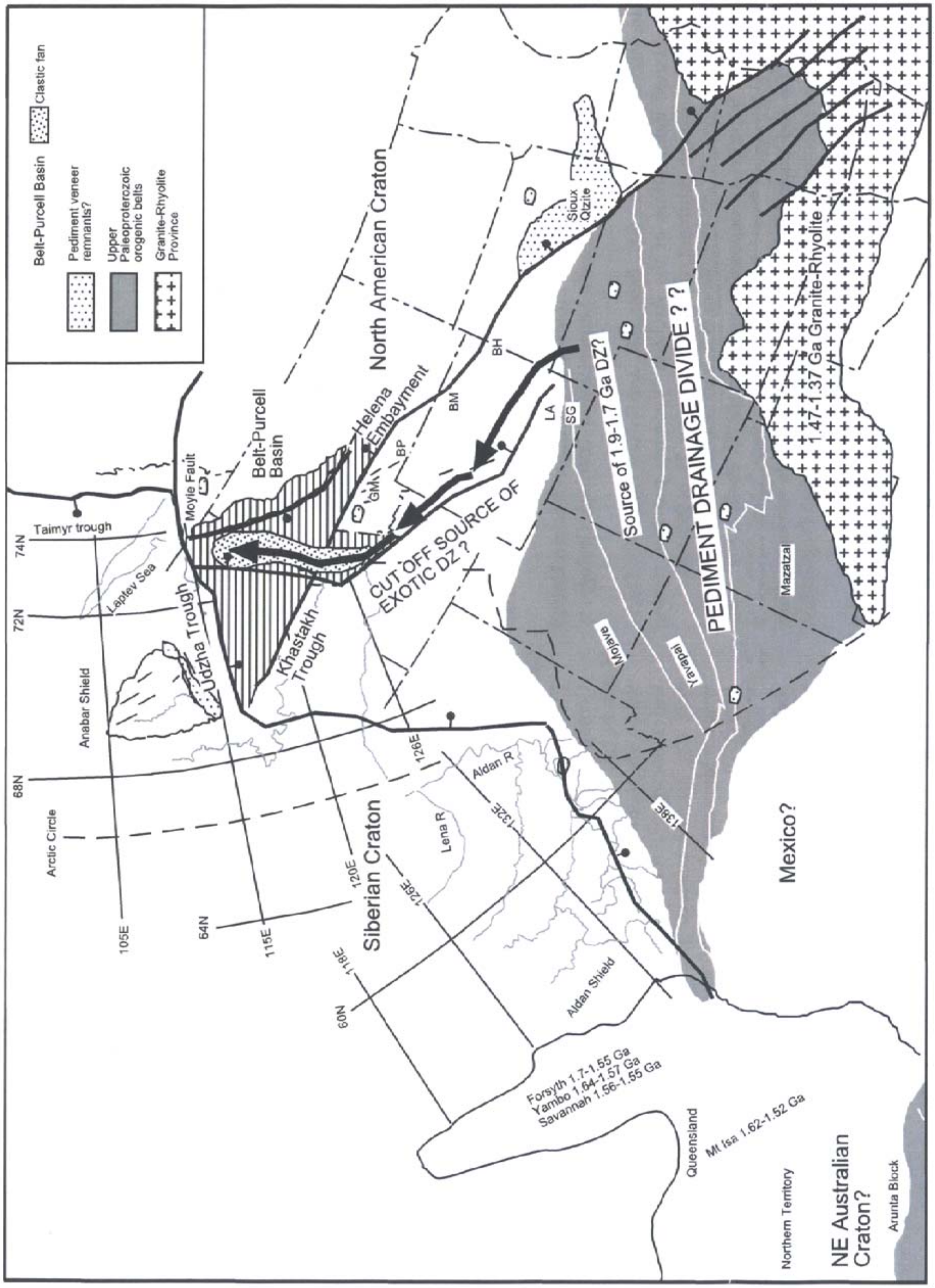


Figure 12. Paleocontinental reconstruction during Missoula Group time shows source areas far to the southeast without significant input from western, non-North American sources that were dominant earlier during deposition of the Lower Belt, Ravalli, and Piegan Groups. From Sears, 2007b.

ester and others, 2016) and removing 2,000 m (1.3 mi) of the lowest strata (Lonn, 2017). The IGS–MBMG team demonstrated the continuity of the Lemhi rocks across the postulated terrane-bounding structures along the Idaho–Montana border into Montana (Lonn and others, 2016a, b; fig. 1). They showed that the Lemhi strata interfinger with and grade laterally northward into the much thinner (<3,500 m; 2.2 mi) and finer-grained Missoula Group (Lonn, 2014; Lonn and others, 2016b; fig. 13). Indeed, U-Pb detrital zircon data show that the Missoula Group and Lemhi strata share the same provenance, including the same upward trend in 1760–1800 Ma grains at higher stratigraphic levels (Link and Fanning, 2003; Link and others, 2007, 2013, 2016; Gardner, 2008; Hendrix and others, 2016; Stewart and others, 2010, 2014; Lonn and others, 2019). If Lemhi Subbasin strata are equivalent to the Missoula Group, subsidence in the Lemhi Subbasin must have been much faster in order to deposit 14 km (8.7 mi) of sediment there versus the 3 km (1.9 mi) thickness of the Missoula Group. Therefore, it appears that the sandy Lemhi strata formed in a southern arm of the Belt Basin, the Lemhi Subbasin, and are relicts of the southeastern, upstream ends of the Missoula Group alluvial aprons (fig. 14). Similarly, the type Lemhi strata may also grade laterally westward into the finer-grained Mesoproterozoic rocks of the Salmon River Mountains (table 1; fig. 14; Burmester and others, 2015; Lonn and others, 2016b). In contrast, Tysdal (2000a, 2003; Tysdal and others, 2005) interpreted the Lemhi succession as marine-tidal, turbidite, and fluvial deposits; he did not speculate on specific correlations with the rest of the Belt.

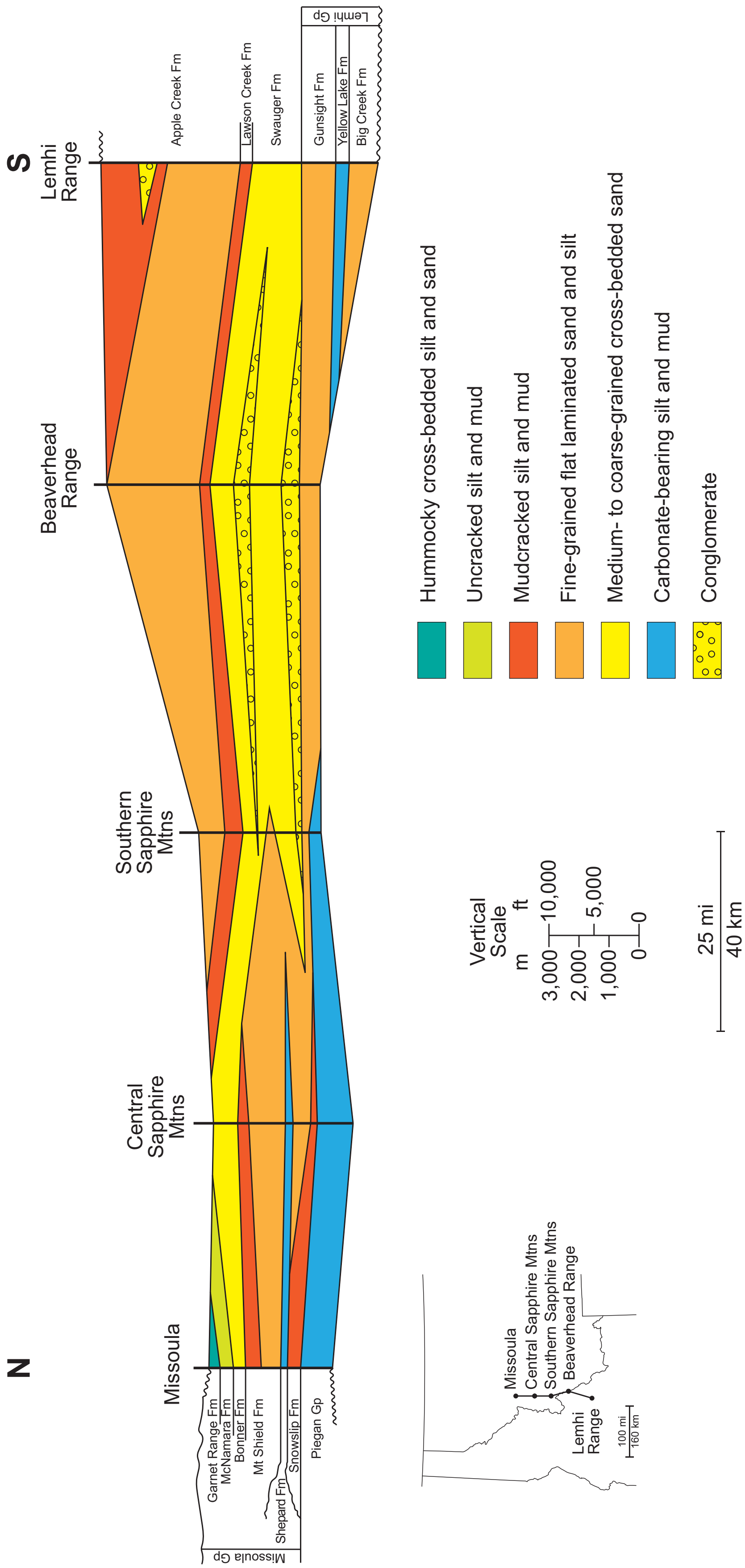


Figure 13. Stratigraphic fence diagram shows facies relationships between Lemhi Subbasin strata and the Missoula Group of the upper Belt Supergroup. Northward, thick sand wedges pinch out and grade into finer-grained and thinner-bedded lithologies as northward-flowing sheetfloods ponded basinward. The bottom of the Lawson Creek Fm correlates with the bottom of the McNamara Fm. From Lonn and others, 2016b.

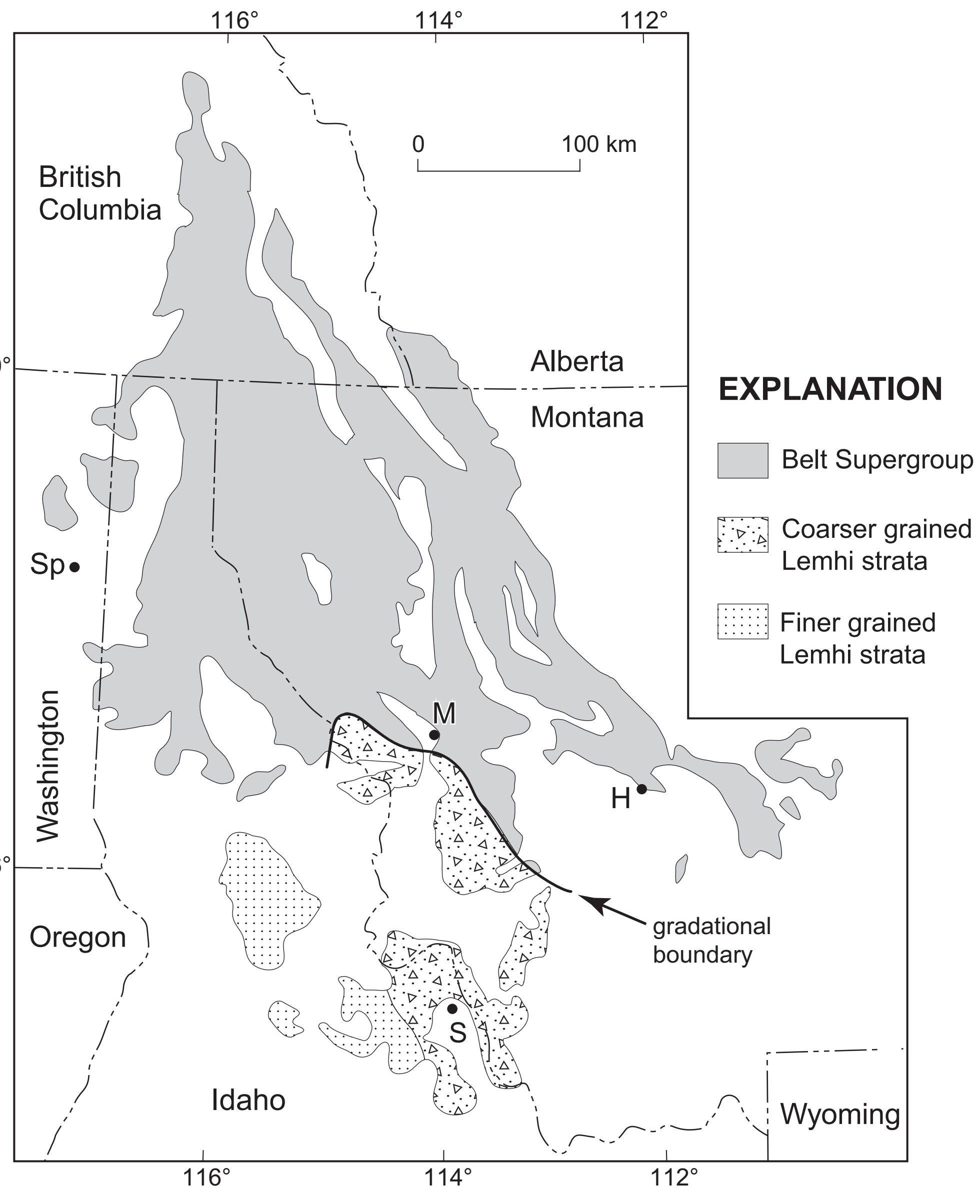


Figure 14. The Lemhi–Missoula Group alluvial apron varied in extent over time, but appears to have originated in modern-day east-central Idaho and southwestern-most Montana. It extended NNW into the main Belt basin, interfingering with and grading into finer-grained sediments of the Missoula Group. Finer-grained Lemhi strata shown to its west are the Yellowjacket, Hoodoo, and western Apple Creek Formations, which probably have similar basinward relationships to the sandier Lemhi strata. H, Helena; M, Missoula; S, Salmon. Modified from Lonn and others, 2016b.

The east side of the Lemhi Subbasin contains the Black Lion Formation of small outcrop extent that appears to be a coarse basin-margin facies similar to the LaHood Formation in the eastern lower Belt (McDonald and others, 2012; McDonald and Lonn, 2013). It appears to grade westward into the Lemhi strata of the upper Belt, and like the much older LaHood Formation, was probably deposited along a fault-bounded margin.

BELT FOSSILS

Fossils have been of considerable interest because they may yield insight into the Mesoproterozoic evolution of life and help in the lacustrine versus marine debate. Walcott (1899) recognized stromatolites in Belt rocks, and they are common from the Piegan Group on up, indicating a low-energy, photic zone (shallow water) environment. Horodyski (1993) reviewed other, possibly eukaryotic, Belt fossils, mostly from the lower Belt, and discounted many of them as non-fossils or the result of prokaryotes influencing sedimentation. The most famous Belt fossil, the "string of beads" from the Appekunny Formation in Glacier National Park, was later named after him (*Horodyskia*). Because similar strings of beads occur in other Mesoproterozoic strata across the world, they were considered to be true multicellular eukaryote fossils (Horodyski, 1993). Retallack and others (2013) later proposed that the beads represented endolichen bladders. However, Rule and Pratt (2017) concluded that the beads resulted from sediment–microbial mat interactions, and that they are really a microbial-induced sedimentary structure.

Recently, Adam and others (2014, 2016) collected microfossils thought to be eukaryotes from the eastern lower Belt of the Helena Embayment. Through comparisons with microfossils in other Mesoproterozoic sediments, they concluded that fossils from the Chamberlain Formation indicate very shallow marine water, in contrast to the Greyson Formation microfossils farther upsection that indicate deeper marine water in a distal shelf environment.

BELT BASIN PALEOGEOGRAPHY AND TECTONICS

The thickness of the Belt Supergroup, the rapid subsidence-sediment accumulation rates, and the mafic composition of syndepositional magmas support the interpretation that the Belt Basin formed in an intracratonic rift within the Columbia/Nuna supercontinent (Winston and Link, 1993; Anderson and Davis, 1995; Ryan and Buckley, 1998; Sears and others, 1998; Evans and others, 2000; Lydon, 2005). Two paleogeographic reconstructions of the basin (figs. 15, 16; Blakey, 2016; Sears, 2007a) both show an elongate north to northwest basin that is largely surrounded by continental crust.

Various workers have described evidence for syndepositional normal faulting within the Belt Basin. Faults have been inferred on the basis of: (1) spatial changes in thicknesses of units and sedimentary facies; (2) soft-sediment deformation that was inferred to occur during syndepositional seismicity; (3) basin-bounding normal faults associated with wedges of coarse conglomerate; and (4) the rapid accumulation of Belt sediments (fig. 2). Lydon (2007) wrote that synsedimentary faults had their most profound effect in the lower Belt where a great contrast in thickness exists between the turbidites and intercalated sills of the Aldridge/Prichard rift-fill sequence (12 km thick; Höy and others, 2000) and the stratigraphically equivalent shelf facies of the Greyson, Newland, Chamberlain, and Neihart Formations (3 km thick; Chandler, 2000). Rapid lateral thickness changes in stratigraphically higher Belt formations, particularly in the sandier units, are also assumed to be the result of syndepositional faults (Winston, 1986a; O'Neill, 1998; Ryan and Buckley, 1998; White and others, 2000; Lydon, 2000, 2005, 2007; Lydon and van Breeman, 2013; Hofmann and others, 2003; Sears, 2007a; Lonn, 2015; Lonn and others, 2016b). Steep basin margins with wedges of coarse conglomerate were probably fault controlled (McMannis, 1963; Winston, 1986a; Tysdal, 2003; McDonald and Lonn, 2013). Figures 16, 17, and 18 show major synsedimentary faults postulated by various Belt workers. Figure 16 is palinspastically restored, whereas figures 17 and 18 are not. Faulting was probably much more complex than these diagrams show, with many cross-cutting syndepositional faults that created subbasins within the main basin (e.g., Ryan and Buckley, 1998; O'Neill, 1998; Höy and others, 2000).

Where is the Western Side of the Basin?

The western side of the basin is thought to have been rifted away during the breakup of the supercontinent Rodinia in the Neoproterozoic to early Cambrian (Price, 2008), and now resides on some other continent. However, this is still an unresolved problem. Antarctica (Dalziel, 1991, 1997; Jones and others, 2015), Australia (Brookfield, 1993; Blewett and others, 1998; Karlstrom and others, 1999, 2001; Burrett and Berry, 2000; Jones and others, 2015), and Siberia (Sears and Price, 2000, 2003; Sears and others, 2004) have all been proposed, but none has been proven conclusively. The most convincing evidence is provided for the Siberian connection (Sears and Price, 2000, 2003; Sears and others, 2004), which is also supported by paleomagnetic data (Hofmann and others, 2003). However, Khudoley and others (2003) reported scant isotopic evidence for a Siberian connection, and Don Winston, upon returning from a long field trip to Siberia, was heard to say: "I didn't see any Belt rocks." Perhaps Wallace (1998) was correct when he suggested the western Basin was subducted, never to be seen again, although most workers agree that significant deformation of Belt rocks did not occur until Mesozoic time. However, the end of Belt sedimentation is thought by some (Anderson and Davis, 1995) to coincide with the Kootenay orogeny (1370–1300 Ma), which also could be related to the 1380–1370 Ma "anorogenic" granites in the Salmon, Idaho area (Doughty and Chamberlain, 1996, 2008; Aleinikoff and others, 2012a). Widespread metamorphism with ages around 1100 Ma also has been documented in the Belt (Aleinikoff and others, 2012b; Nesheim and others, 2012; Zirakparvar and others, 2010; Jeff Vervoort,

Figure 15. Blakey's (2016) reconstruction of the North American continent at 1450 Ma (early to middle Belt time) shows the Belt Sea occupying an elongate north–northwest basin open to the ocean on its northern end. From Blakey, https://deeptimemaps.com, with permission (Blakey, 2016).

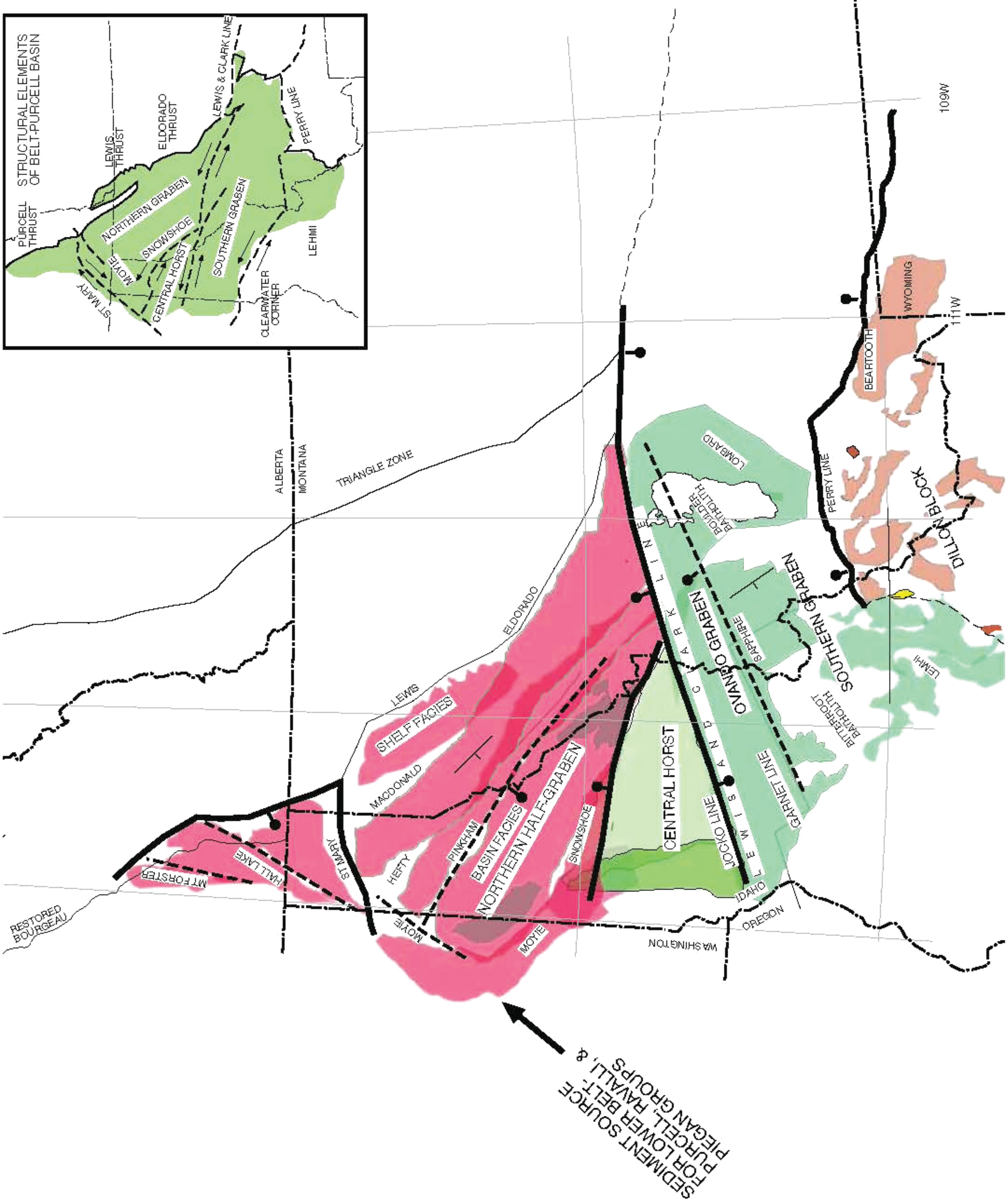


Figure 16. Sears' (2007a) palinspastic reconstruction of the Belt Basin shows an elongate north–north-west basin. Also shown are major postulated syndepositional faults in their restored locations. From Sears, 2007a.

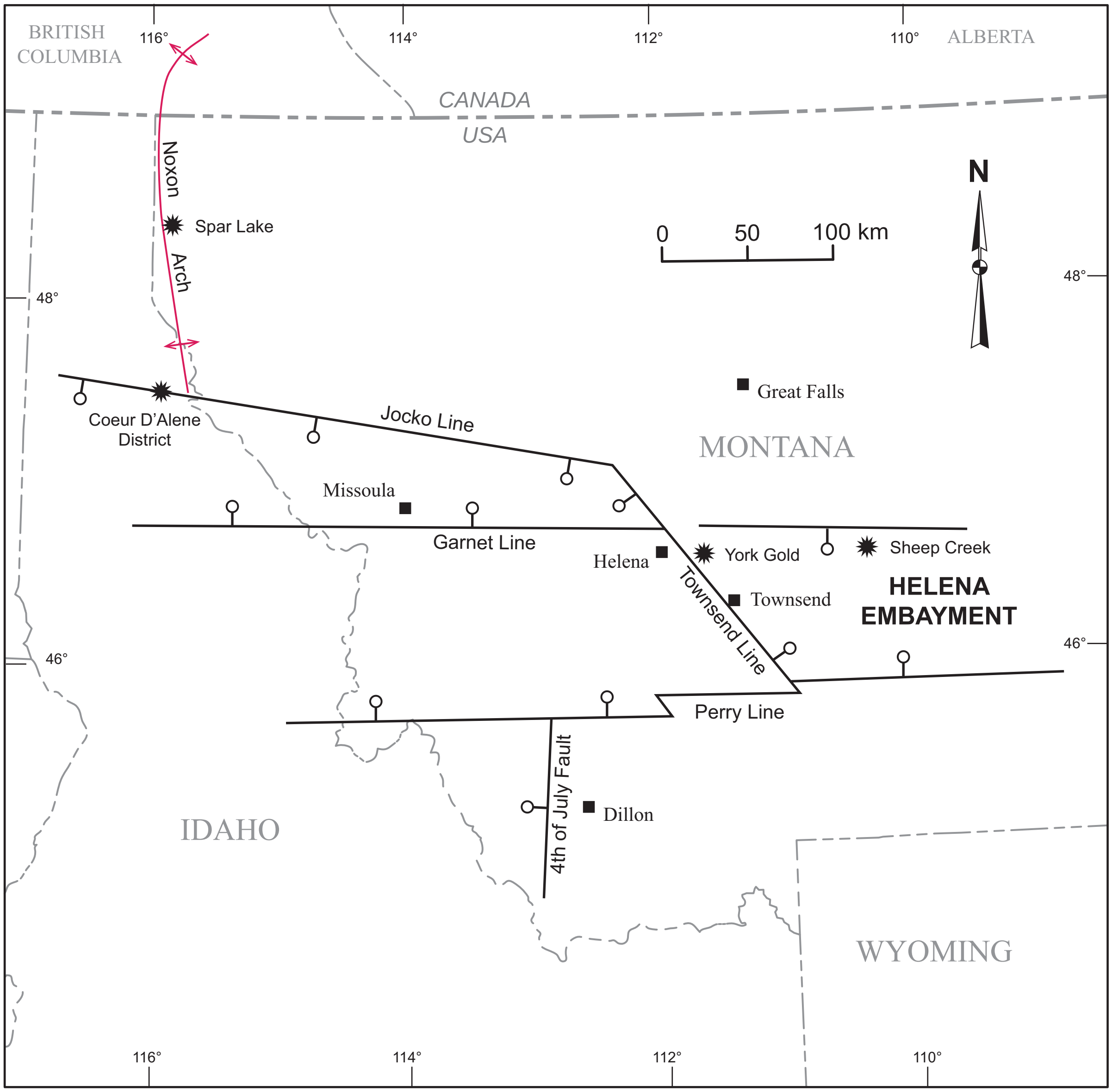


Figure 17. Winston's (1986a) syndepositional growth faults are based on thickness changes and on locations of Phanerozoic structures that were greatly influenced by the Proterozoic structures. Noxon Arch is added from White (2000), White and Appelgate (1999, 2000), and White and others (2000). The 4th of July fault is from McDonald and Lonn (2013). Major stratabound mineral deposits are shown by stars. Modified from Winston, 1986a.

written commun., 2014). This may have resulted from a protracted tectonic event, underplating of the continental crust, or simply recirculating hydrothermal brines.

BELT MINERAL DEPOSITS

Belt rocks host some of the world's most important metallic mineral deposits, including the Coeur d'Alene silver–lead–zinc (Idaho), the Sullivan massive sulfide (British Columbia), the Spar Lake copper–silver (Montana), the Golden Sunlight gold (Montana), and the Blackbird cobalt (Idaho) deposits. Other large deposits awaiting development are the Black Butte copper–lead–zinc deposit (Montana) and the York gold occurrence (Montana). Because these large deposits are stratiform or stratabound, they are thought to have Proterozoic syngenetic or diagenetic origins, and are postulated to be controlled by the many rift-related growth faults discussed above. Later tectonic events remobilized and concentrated the metals in several deposits. See the following papers for more detailed discussion of the Proterozoic origins of these deposits: **Coeur d'Alene:** White and Appelgate, 1999, 2000; White, 2000; Mauk and White, 2004; **Sullivan:** Lydon and others, 2000; **Spar Lake:** Boleneus and others, 2006; **Golden Sunlight:** Foster and others, 1993; **Blackbird:** Bookstrom and others, 2016; **Black Butte**: Graham and others, 2012; Zieg and others, 2013; **York:** Thorson, 1993; Whipple and Morrison, 1993. More information on Belt mineral deposits can

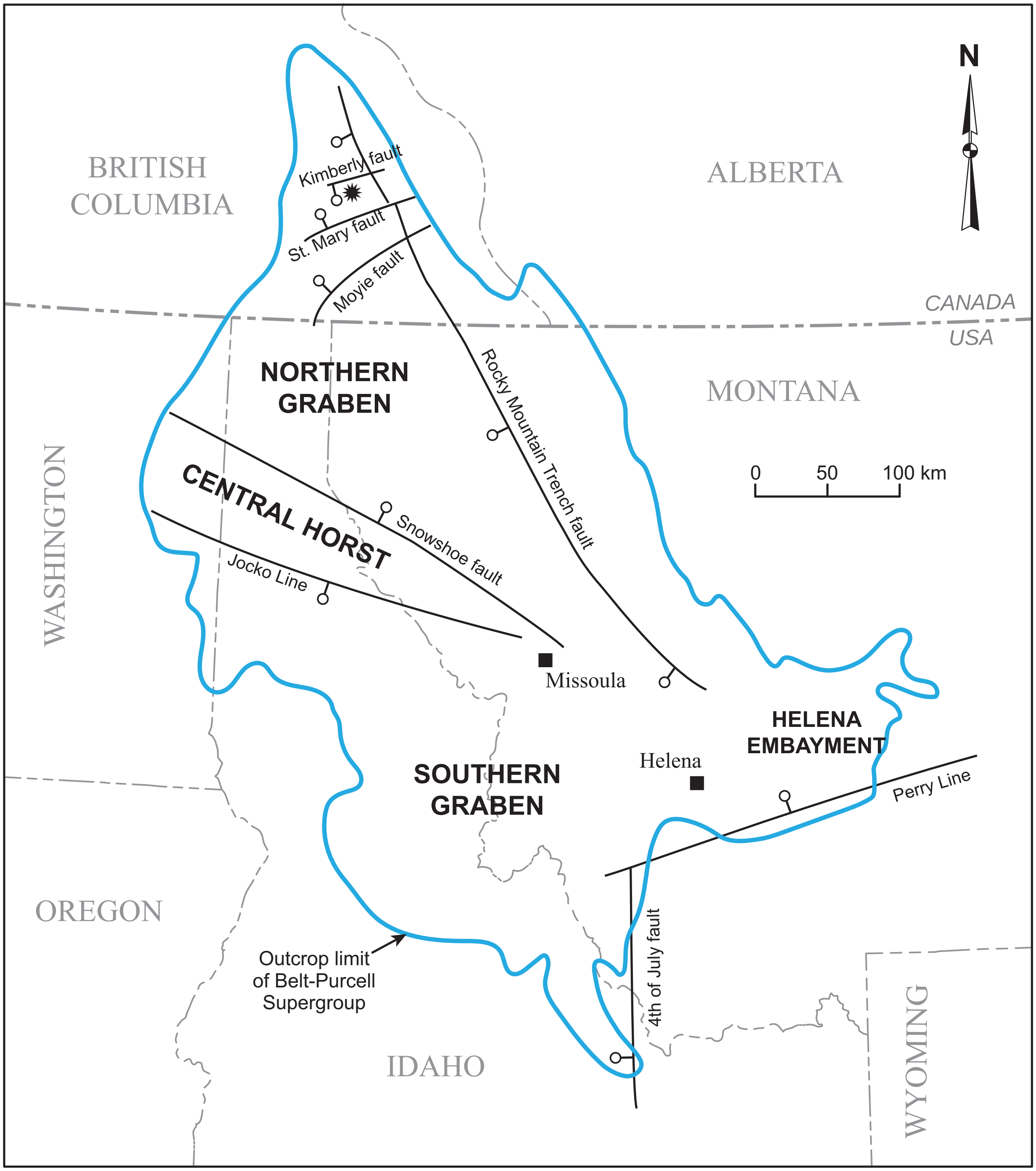


Figure 18. Lydon's interpretation of synsedimentary growth faults in the Belt Basin is based on thicknesses of the lower Belt in the northern part of the Belt Basin and on Sears' (2007a) unrestored syndepositional faults (see fig. 16, this paper). Star marks location of the Sullivan deposit along the Kimberly fault. Compiled from Lydon (2005, 2007) and Lydon and van Breemen (2013).

be found in this volume (Gammons and others, 2020), in Lydon and other's (2000) summary of the Sullivan deposit, and in a special issue of *Economic Geology* devoted to the origins of mineral deposits in Belt strata (see Box and others, 2012).

WAS THE BELT SEA A LAKE OR PART OF THE OCEAN?

The marine versus lacustrine question has generated lively discussion among geologists for decades. Detailed discussion has been presented above for each group, and is summarized here. Detrital zircon data and facies relationships show that the Belt Sea had sediment sources to the east, south, and west, and so was mostly surrounded by land throughout its history.

The turbidites and deltaic sediment of the lower Belt could have been deposited in either an ocean or a large, deep lake or sea. Isotopic evidence (Lyons and others, 1998, 2000; Anderson and Davis, 1995; Luepke and Lyons, 2001) favors a restricted marine setting with episodic incursions of seawater. Perhaps the paleogeography during lower Belt time matched that shown in figure 15 (Blakey, 2016), where the basin was open to the ocean at least periodically on the north or northwest.

Beginning with the Ravalli Group, though, Belt rocks largely represent shallow water deposits. According to Winston (2016), the Ravalli Group is composed largely of fluvial to shallow standing water strata deposited on immense sand and mud flats. Winston cites the lack of tidal features, especially channels, in both the Ravalli Group and overlying Piegan Group as evidence of a lacustrine environment (Winston, 1986c, 1998, 2007, 2016). Lyons and others (1998) provided isotopic evidence that the Piegan Group was lacustrine. The hummocky crossbeds that characterize the Piegan Group were likely formed in shallow, storm-tossed seas, and are probably best explained using the lacustrine model. However, they could have been generated in a restricted marine environment or in shallow marine waters at the edge of the ocean (Pratt 2001, 2017a, b; Johnson, 2013; Frank and others, 1997), and the ocean waters could have more easily replenished the calcium, magnesium, and carbonate ions abundant in the Piegan Group (Link, 1998).

Perhaps the well-studied Missoula Group, with its great thickness of mudcracked deposits, contains the best strata to debate. It is hard to explain the greater than 3 km (1.9 mi) thickness of the Missoula Group's mudcracked sediment as a marine deposit where subsidence exactly matched the rate of sedimentation for millions of years. The playa lake model of Winston probably better explains this persistently shallow water, the repeated drying of the mudflats, and the sheetfloods that crossed the surrounding alluvial plains without incising channels. In such a closed basin, accumulation of sediments consistently increased the base and grade levels, explaining the lack of incision and the persistence of shallow water.

Still, there is some evidence for the entire Belt to be marine (Pratt, 2001). Horodyski (1993) cited the similarity of Belt microfossils to Mesoproterozoic microfossils worldwide to indicate all are derived from a marine environment. Schieber (1998) used chemical mass balance to compare Canadian Shield crust to Belt geochemistry and supported a marine interpretation because of the influx and export of some chemicals. Schieber (1998) also cited reports of glauconite throughout the Belt section as evidence of deep marine conditions, but glauconite has also been reported from saline lagoonal environments (Albani and others, 2005; Suttill, 2009). Although the lake versus ocean debate may never be settled, there appears to be movement toward acceptance of the lacustrine model for much of the Belt.

SUMMARY

Rifting within the supercontinent Columbia/Nuna about 1.48 Ga formed a huge north–northwest-elongate basin open to the ocean only on its northwestern end. A Mississippi River-scale delta system on its southwestern side shed turbidites into the rapidly deepening basin, depositing as much as 12 km (7.5 mi) of sediment as the lower Belt "group." As the basin became more and more isolated from the ocean, it filled and shallowed. An enormous alluvial apron complex prograded from the west and southwest, built by repeated sheetfloods that ponded against the tectonically quiescent Laurentian land mass to the east, depositing the Ravalli Group. More rifting caused perennial, but shallow, lake complexes to span the Belt Basin, expanding westward through Piegan Group time. The tectonic configuration changed at the beginning of Missoula Group time, with subsidence waning in the main basin and increasing in the Lemhi Subbasin on the south end. The western sediment source was cut off and enormous alluvial megafans advanced across the lake basin from the south and southeast. They deposited more than 14 km (2.5 mi) of feldspathic quartzite in the Lemhi Subbasin near their heads and the thinner and finer-grained Missoula Group strata nearer to the center of the basin. By the time the record of Belt deposition ended about 1380 Ma, as much as 18 km (11 mi) of sediment had accumulated in the basin.

It is evident that the Belt Basin was like no intracratonic basin on earth today. Its featureless and desolate alluvial plains stretched beyond the horizons. The occasional sheetfloods rolling down its megafans carried massive volumes of sediment-laden water, and its saline seas were of enormous extent but mostly very shallow and muddy. The sedimentary succession that resulted is the thickest on earth. Although much progress has been made in understanding the alien landscape that was the Belt Basin, many questions remain and await resolution by those geologists seeking a lifetime of challenging work.

REFERENCES CITED

Adam, Z.R., Mogk, D., Skidmore, M.L., and Butterfield, N.J., 2014, Microfossils from the Greyson Formation, Lower Belt Supergroup: Support for early Mesoproterozoic biozonation: Geological Society of America Abstracts with Programs, v. 46, no. 5, p. 71.

Adam, Z.R., Skidmore, M.L., and Mogk, D.W., 2016, Paleoenvironmental implications of an expanded microfossil assemblage from the Chamberlain Formation, Belt Supergroup, Montana, *in* MacLean, J.S., and Sears, J.W., eds., Belt Basin: Window to Mesoproterozoic Earth: Geological Society of America Special Paper 522, Chapter 4, p. 101–119, doi: https://doi.org/10.1130/2016.2522(04)

Albani, A.E., Meunier, A., and Fursich, F., 2005, Unusual occurrence of glauconite in a shallow lagoonal environment (Lower Cretaceous, northern Aquitaine Basin, SW France): Terra Nova, v. 17, no. 6, p. 537–544, doi: https://doi.org/10.1111/j.1365-3121.2005.00646.x

Aleinikoff, J.N., Slack, J.F., Kund, K., Evans, K.V., Fanning, C.M., Mazdab, F.K., Wooden, J.I., and Pillers, R.M., 2012a, Constraints on the timing of Co-Cu+/-Au mineralization in the Blackbird district, Idaho, using SHRIMP U-Pb age of monazite and xenotime plus zircon ages of related Mesoproterozoic orthogneisses and metasedimentary rocks: Economic Geology, v. 107, no. 6, p. 1143–1175, doi: https://doi.org/10.2113/econgeo.107.6.1143

Aleinikoff, J.N., Hayes, T.S., Evans, K.V., Mazdab, F.K., Pillers, R.M., and Fanning, C.M., 2012b, SHRIMP U–Pb ages of xenotime and monazite from the Spar Lake red bed-associated Cu–Ag deposit, western Montana: Implications for ore genesis: Economic Geology, v. 107, p. 1251–1274, doi: https://doi.org/10.2113/econgeo.107.6.1251

Anderson, H.E., and Davis, D.W., 1995, U-Pb geochronology of the Moyie sills, Purcell Supergroup, southeastern British Columbia: Implications for the Mesoproterozoic geologic history of the Purcell (Belt) Basin: Canadian Journal of Earth Sciences, v. 32, p. 1180–1193, doi: https://doi.org/10.1139/e95-097

Baldwin, J.A., Lewis R.S., Vervoort, J.D., and McDonie, C.D., 2016, Metamorphic history of the Belt Supergroup and underlying Paleoproterozoic basement rocks in the western part of the Clearwater complex, Idaho, *in* Lewis, R.S., and Schmidt, K.L., eds., Exploring the geology of the inland northwest: Geological Society of America Field Guide 41, p. 251–264, doi: https://doi.org/10.1130/2016.0041(08)

Balgord, E., Forgette, M., Mahoney, J.B., Ihinger, P., Kimbrough, D., 2008, Redefinition of the Precambrian-Cambrian contact in southwestern Montana: Geological Society of America Abstracts with Programs, v. 40, no. 6, p. 167.

Barrell, J., 1906, Relative geological importance of continental, littoral, and marine sedimentation, Part II: Journal of Geology, v. 14, p. 430-457.

Binda, P.L., and Koopman, H.T., 1998, The Empire Formation in southwestern Alberta and southeastern British Columbia, Canada, *in* Berg, R., ed., Proceedings of Belt Symposium III: Montana Bureau of Mines and Geology Special Paper 112, p. 56-69.

Bishop, J.W., Sumner, D.Y., and Huerta, N.J., 2006, Molar-tooth structures of the Neoarchean Monteville Formation, Transvaal Supergroup, South Africa: Sedimentology, v. 53, p. 1049-1082, doi: https://doi.org/10.1111/j.1365-3091.2006.00802.x

Blakey, R.C., 2016, Paleogeographic maps, Colorado Plateau Geosystems Inc.

Blewett, R.S., Black, L.P., Sun, S.S., Knutson, J., Hutton, L.J., and Bain, J.H.C., 1998, U–Pb zircon and Sm–Nd geochronology of the Mesoproterozoic of North Queensland: Implications for a Rodinian connection with the Belt Supergroup of North America: Precambrian Research, v. 89, p. 101–127, doi: https://doi.org/10.1016/S0301-9268(98)00030-8

Boleneus, D., Appelgate, L.M., Zientek, M.L., Carlson, M.H., Assmus, K.C., and Chase, D.W., 2006, Stratigraphic control of copper-silver deposits in the Mesoproterozoic Revett Formation, Idaho and Montana: Northwest Geology, v. 35, p. 57–60.

Bookstrom, A.A., Box, S.E., Cossette, P.M., Frost, T.P., Gillerman, V.S., King, G.R., and Zirakparvar, N.A., 2016, Geologic history of the Blackbird Co-Cu district in the Lemhi Subbasin of the Belt-Purcell Basin, *in* MacLean, J.S., and Sears, J.W., eds., Belt Basin: Window to Mesoproterozoic Earth: Geological Society of America Special Paper 522, Chapter 5, p. 185–219, https://doi.org/doi: 10.1130/2016.2522(08)

Box, S.E., Bookstrom, A.A., and Anderson, R.G., 2012, Origins of mineral deposits, Belt-Purcell Basin, United States and Canada: An introduction: Economic Geology, v. 107, no. 6, p. 1081–1088, doi: https://doi.org/10.2113/econgeo.107.6.1081

Breckenridge, R.M., Burmester, R.F., Lewis, R.S., and McFaddan, M.D., 2014, Geologic map of the east half of the Bonners Ferry 30 x 60 minute quadrangle, Idaho and Montana: Idaho Geologic Survey Digital Web Map, DWM-173, scale 1:75,000.

Brookfield, M.E., 1993, Neoproterozoic Laurentia-Australia fit: Geology, v. 21, p. 683–686, doi: https://doi.org/10.1130/0091-7613(1993)021%3C0683:NLAF%3E2.3.CO;2

Browne, J.L., 2006, Geologic map of the Cataldo quadrangle, Kootenai and Shoshone Counties, Idaho: Idaho Geological Survey Technical Report T-06-5, scale 1:24,000.

Browne, J.L., 2017, Geologic map of the Kellogg East and Southern Grizzly Mountain quadrangles, Shoshone County, Idaho: Idaho Geological Survey Technical Report T-17-1, scale 1:24,000.

Buckley, S.N., and Sears, J.W., 1998, Emplacement of mafic sills into wet sediments in the Prichard Formation, Middle Proterozoic Belt Supergroup, Perma area, western Montana, *in* Berg, R., ed., Proceedings of Belt Symposium III: Montana Bureau of Mines and Geology Special Paper 112, p. 32–43.

Buddington, A.M., Wang, D., and Doughty, P.T., 2016, Pre-Belt basement tour: Late Archean-Early Proterozoic rocks of the Cougar Gulch area, southern Priest River complex, Idaho, *in* Lewis, R.S., and Schmidt, K.L., eds., Exploring the geology of the Pacific Northwest: Geological Society of America Field Guide 41, p. 265–284.

Burmester, R.F., Lewis, R.S., and Bennett, E.H., 1998, Belt rocks north of the Bitterroot lobe of the Idaho batholith: The southwest basin margin?, *in* Berg, R., ed., Proceedings of Belt Symposium III: Montana Bureau of Mines and Geology Special Paper 112, p. 145–156.

Burmester, R.F., Lonn, J.D., Lewis, R.S., and McFaddan, M.D., 2013, Toward a grand unified theory for stratigraphy of the Lemhi Subbasin of the Belt Supergroup: Northwest Geology, v. 42, p. 1–20.

Burmester, R.F., Lewis, R.S., Lonn, J.D., and McFaddan, 2015, The Hoodoo is the Swauger and other heresies: Lemhi Subbasin correlations and structures, east-central Idaho: Northwest Geology, v. 44, p. 73–88.

Burmester, R.F., Lonn, J.D., Lewis, R.S., and McFaddan, M.D., 2016, Stratigraphy of the Lemhi Subbasin of the Belt Supergroup, *in* MacLean, J.S., and Sears, J.W., eds., Belt Basin: Window to Mesoproterozoic Earth: Geological Society of America Special Paper 522, Chapter 5, p. 121–138, doi: https://doi.org/10.1130/2016.2522(05)

Burrett, C., and Berry, R., 2000, Proterozoic Australia-western United States (AUSWUS) fit between Laurentia and Australia: Geology, v. 28, p. 103–106, doi: https://doi.org/10.1130/0091-7613(2000)28%3C103:PAUSAF%3E2.0.CO;2

Calkins, F.C., and Emmons, W.H., 1915, Geologic atlas of the United States: U.S. Geological Survey Folio GF-196, scale 1:125,000.

Campbell, A.B., 1960, Geology and mineral deposits of the St. Regis-Superior area, Mineral County, Montana: U.S. Geological Survey Bulletin 1082-I, 612 p.

Chandler, F.W., 2000, The Belt-Purcell Basin as a low-latitude passive rift: Implications for the geological environment of Sullivan-type deposits, *in* Lydon, J.W., Höy, T., Slack, J.F., and Knapp, M.E., eds., The geological environment of the Sullivan deposit, British Columbia: Geological Association of Canada, Mineral Deposits Division (MDD) Special Publication no. 1, p. 82–112.

Collins, B.W., and Winston, D., 1999, The intrastratal middle Wallace breccia: Evidence of post-Wallace faulting: Geological Society of America Abstracts with Programs, v. 31, no. 4, p. 8.

Constenius, K.N., Velencia, V., Ibanez-Mejia, M., and Domanik, K.J., 2017, Peperite in the Purcell Lava and a revised age of the upper Proterozoic Belt-Purcell Supergroup: Geological Society of

America Abstracts with Programs, v. 49, no. 5, doi: https://doi.org/10.1130/abs/2017RM292776

Cook, F.A., and van der Velden, A.G., 1995, Three dimensional crustal structure of the Purcell anticlinorium in the Cordillera of southwestern Canada: Geological Society of America Bulletin, v. 107, p. 642–664, doi: https://doi.org/10.1130/0016-7606(1995)107%3C0642:TDCSOT%3E2.3.CO;2

Cressman, E.R., 1985, Lower part of the Belt Supergroup (Middle Proterozoic), near Plains, Sanders County, Montana: U.S. Geological Survey Bulletin 1553, 70 p.

Cressman, E.R., 1989, Reconnaissance stratigraphy of the Prichard Formation (Middle Proterozoic) and the early development of the Belt Basin, Washington, Idaho, and Montana: U.S. Geological Survey Professional Paper 1490, 80 p.

Dalziel, I.W.D., 1991, Pacific margins of Laurentia and East Antartica-Australia as a conjugate rift pair: Evidence and implications for an Eocambrian supercontinent: Geology, v. 19, p. 598–601, doi: https://doi.org/10.1130/0091-7613(1991)019%3C0598:PMOLAE%3E2.3.CO;2

Dalziel, I.W.D., 1997, Overview: Neoproterozoic-Paleozoic geography and tectonics: Review, hypothesis, environmental speculations: Geological Society of America Bulletin, v. 109, p. 16-42, doi: https://doi.org/10.1130/0016-7606(1997)109%3C0016:ONPGAT%3E2.3.CO;2

Doughty, P.T., and Chamberlain, K.R., 1996, Salmon River arch revisited: New evidence for 1370 rifting near the end of deposition in the Middle Proterozoic Belt Basin: Canadian Journal of Earth Sciences, v. 33, p. 1037–1052, doi: https://doi.org/10.1139/e96-079

Doughty, P.T., and Chamberlain, K.R., 2008, Protolith age and timing of Precambrian magmatic and metamorphic events in the Priest River Complex, northern Rockies: Canadian Journal of Earth Sciences, v. 45, p. 99–116, doi: https://doi.org/10.1139/cjes-35-1-39

Elston, D.P., Enkin, R.J., Baker, J., and Kisilevsky, D.K., 2002, Tightening the Belt: Paleomagnetic-stratigraphic constraints on deposition, correlation, and deformation of the Middle Proterozoic (ca. 1.4 Ga) Belt-Purcell Supergroup, United States and Canada: Geological Society of America Bulletin, v. 114, p. 619–638, doi: https://doi.org/10.1130/0016-7606(2002)114<0619:TTBPSC>2.0.CO;2

Evans, K.V., Aleinikoff, J.N., Obradovich, J.D., and Fanning, C.M., 2000, SHRIMP U-Pb geochronology of volcanic rocks, Belt Supergroup, western Montana: Evidence for rapid deposition of sedimentary strata: Canadian Journal of Earth Sciences, v. 37, p. 1287–1300, doi: https://doi.org/10.1139/cjes-37-9-1287

Evans, K.V., and Green, G.G., compilers, 2003, Geologic map of the Salmon National Forest and vicinity, east-central Idaho: U.S. Geological Survey, Geologic Investigations Series Map I-2765, 2 sheets, scale 1:100,000.

Farooqui, M.A., 1997, Provenance of the Bonner Formation (Middle Proterozoic Belt Supergroup), Montana, and the Buffalo Hump Formation (Deer Trail Group), Washington: Petrographic evidences for multiple source terrains, *in* Berg, R.B., ed., Belt Symposium III Abstracts: Montana Bureau of Mines and Geology Special Publication 381, p. 17–19.

Foster, F., Chadwick, T., and Nilsen, T.H., 1993, Paleodepositional setting and synsedimentary mineralization in Belt Supergroup rocks of the Whitehall, Montana, area, *in* Berg, R.B., compiler, Belt Symposium III Abstracts: Montana Bureau of Mines and Geology Open-File Report 381, p. 23–25.

Frank, T.D., Lyons, T.W., and Lohman, K.C., 1997, Isotopic evidence for the paleoenvironmental evolution of the Mesoproterozoic Helena Formation, Belt Supergroup, Montana, USA: Geochemica et Cosmochimica Acta, v. 61, p. 5023–5041, doi: https://doi.org/10.1016/S0016-7037(97)80341-9

Furniss, G., Rittel, J.F., and Winston, D., 1998, Gas bubble and expansion crack origin of “molar-tooth” calcite structures in the Middle Proterozoic Belt Supergroup, western Montana: Journal of Sedimentary Research, v. 68, p. 104–114, doi: https://doi.org/10.2110/jsr.68.104

Gammons, C.H., Korzeb, S., and Hargrave, P., 2020, Metallic ore deposits of Montana, *in* Metesh, J.J., and Gammons, C.H., eds., Geology of Montana—Special Topics: Montana Bureau of Mines and Geology Special Publication 122: v. 2, 30 p.

Gardner, D.W., 2008, Sedimentology, stratigraphy, and provenance of the Upper Purcell Supergroup, southeastern British Columbia, Canada: Implications for syn-depositional tectonism, basin models, and paleogeographic reconstructions: Victoria, British Columbia, University of Victoria, M.S. thesis, 85 p.

Gellatly, A.M., and Winston, D., 1998, Is "molar-tooth" calcite in the Helena Formation (Middle Proterozoic Belt Supergroup, Montana) inverted vaterite?: Geological Society of America Abstracts with Programs, v. 30, no. 7, p. A333.

Gellatly, A.M., and Lyons, T.W., 2005, Trace sulfate in mid-Proterozoic carbonates and the sulfur isotope record of biospheric evolution: Geochimica et Cosmochimica Acta, v. 69, p. 3813–3829, doi: https://doi.org/10.1016/j.gca.2005.01.019

Glombick, P., Brown, D.A., and MacLeod, R.F., compilers, 2010a, Geology, Creston, British Columbia: Geological Survey of Canada Open File 6152, scale 1:50,000.

Glombick, P., Brown, D.A., and MacLeod, R.F., compilers, 2010b, Geology, Yahk, British Columbia: Geological Survey of Canada Open File 6153, scale 1:50,000.

Godard, E.N.J., 1998, Sediment dispersal pathways in quartzite units of the lower Prichard Formation, Pine Creek drainage, Idaho, *in* Berg, R., ed., Proceedings of Belt Symposium III: Montana Bureau of Mines and Geology Special Paper 112, p. 44–69.

Godlewski, D.W., 1981, Origin and classification of the middle Wallace breccias: Missoula, University of Montana, M.S. thesis, 74 p.

Graham, G., Hitzman, M.W., and Zieg, J., 2012, Geologic setting, sedimentary architecture, and paragenesis of the Mesoproterozoic sediment-hosted Sheep Creek Cu-Co-Ag deposit, Helena Embayment, Montana: Economic Geology, v. 107, no. 6, p. 1115–1141, doi: https://doi.org/10.2113/econgeo.107.6.1115

Griggs, A.B., 1973, Geologic map of the Spokane quadrangle, Washington, Idaho, and Montana: U.S. Geological Society Miscellaneous Geologic Investigations Map I-768, scale 1:250,000.

Guerrerro, J.C., Mueller, P.A., and Mogk, D.W., 2016, Provenance study and geochemical analysis of the LaHood Formation of the Bridger Range, Montana: Northwest Geology, v. 45, p. 21–28.

Harazim, D., Callow, R.T.H., and McIlory, D., 2013, Microbial mats implicated in the generation of intrastratal shrinkage (synearesis) cracks: Sedimentology, v. 60, p. 1621–1638, doi: https://doi.org/10.1111/sed.12044

Harrison, J.E., Griggs, A.B., and Wells, J.D., 1986, Geologic and structure maps of the Wallace 1° x 2° quadrangle, Montana and Idaho: U.S. Geological Survey Miscellaneous Investigations Series Map I-1509-A, scale 1:250,000.

Harrison, J.E., Cressman, E.R., and Whipple, J.W., 1992, Geologic map of the Kalispell 1° x 2° quadrangle, Idaho and Montana: U.S. Geological Survey Miscellaneous Investigations Series Map I-2267, scale 1:250,000.

Harrison, J.E., Whipple, J.W., and Kidder, D.L., 1997, Belt Supergroup stratigraphy and structure, north-central Belt Basin, northwestern Montana, *in* Link, P.K., ed., Geologic Guidebook to the Belt-Purcell Supergroup, Glacier National Park and vicinity, Montana and adjacent Canada: Belt Symposium III, Belt Association, Spokane, Washington, p. 1–19.

Hendrix, M.S., Winston, D., Crowley, J.L., and Schmitz, M.D., 2016, Detrital zircon results from Proterozoic and Paleozoic rocks of western Montana and implications for geologic evolution of the western North American margin: Geological Society of America Abstracts with Programs 48, no. 6, doi: https://doi.org/10.1130/abs/2016RM-276264.

Hofmann, M.H., Mocko, J., and Yalon, N., 2003, The missing part of the Belt Basin: Re-evaluating paleomagnetic and geologic evidences of different paleogeographic reconstructions: Northwest Geology, v. 32, p. 199–202.

Horodyski, R.J., 1993, Precambrian paleontology of the western conterminous U.S. and northwestern Mexico, *in* Reed, J.C., Jr., Bickford, M.E., Houston, R.S., Link, P.K., Rankin, D.W., Sims, P.K., and Van Schmus, W.R., eds., Precambrian: Conterminous U.S.: Geological Society of America Decade of North American Geology, v. C-2, p. 558–565.

Höy, T., 1982, The Purcell Supergroup in southeastern British Columbia: Sedimentation, tectonics, and stratiform lead–zinc deposits, *in* Hutchinson, R.W., Spence, C.W., and Franklin, J.M., eds., Precambrian sulfide deposits: Geological Association of Canada Special Paper 25, p. 127–147.

Höy, T., 1993, Geology of the Purcell Supergroup in the Fernie Westhalf map area, southeastern British Columbia: Geological Survey of Canada Bulletin 84, 157 p.

Höy, T., Price, R.A., Legun, A., Grant, B., and Brown, D., 1995, Purcell Supergroup, southeastern British Columbia, Geological compilation map: British Columbia Geological Survey Branch, Geoscience Map 1995-1, scale 1:250,000.

Höy, T., Anderson, D., Turner, R.J.W., and Leitch, C.H.B., 2000, Tectonic, magmatic, and metallogenic history of the early synrift phase of the Purcell Basin, southeastern British Columbia, Chapter 4, *in* Lydon, J.W., Höy, T., Slack, J.F., and Knapp, M.E., eds., The geological environment of the Sullivan deposit, British Columbia: Geological Association of Canada Mineral Deposits Division (MDD) Special Publication 1, p. 32–60.

Johnson, D.J., 2013, A fining and thinning upward sequence in the Helena and Wallace Formations, mid-Proterozoic Belt Supergroup, Montana and Idaho: Northwest Geology, v. 42, p. 21–38.

Jones, J.V., Daniel, C.G., and Does, M.F., 2015, Purcell Basin and southwestern Laurentia during the Mesoproterozoic, ca. 1.60–1.40 Ga: Lithosphere, v. 7, no. 4, p. 465–472, doi: https://doi.org/10.1130/L438.1

Karlstrom, K.E., Harlan, S.S., Williams, M.L., McLelland, J., Geissman, J.W., and Ahall, K.I., 1999, Refining Rodinia: Geologic evidence for the Australian-Western US connection in the Proterozoic: GSA Today, v. 9, p. 1–7.

Karlstrom, K.E., Ahall, K.I., Harlan, S.S., Williams, M.L., McLelland, J., and Geissman, J.W., 2001, Long-lived (1.8–1.0 Ga) convergent orogeny in southern Laurentia: Its extensions to Australia and Baltica, and implications for refining Rodinia: Precambrian Research, v. 111, p. 5–30, doi: https://doi.org/10.1016/S0301-9268(01)00154-1

Khudoley, A.K., Kropachev, A.P., Rublev, A.G., Lyahnitskaya, O.Y., and Podkovyrov, V.N., 2003, Isotopic data from the Precambrian of the Siberian craton: An overview and implications for paleocontinental reconstructions: Northwest Geology, v. 32, p. 203–204.

Kidder, D.L., 1998, The Libby Formation and its significance with respect to changing sedimentation patterns in the upper Belt Supergroup, *in* Berg, R., ed., Proceedings of Belt Symposium III: Montana Bureau of Mines and Geology Special Paper 112, p. 104–112.

Kříbek, B., Hladíková, J., and Frýda, J., 1997, Scapolite- and anhydrite-bearing rocks from the Moldanubian Zone of the Bohemian Massif: Metamorphosed exhalites and evaporates: Journal of the Czech Geological Society, v. 42 (1997), no. 3, p. 62, available at http://www.jgeosci.org/content/JCGS1997_3__kribek.pdf.

Lemoine, S.R., and Winston, D., 1986, Correlation of the Snowslip and Shepard Formations of the Cabinet Mountains with upper Wallace rocks of the Coeur d'Alene Mountains, western Montana, *in* Roberts, S.M., ed., Belt Supergroup: A guide to Proterozoic rocks of western Montana and adjacent areas: Montana Bureau of Mines and Geology Special Publication 94, p. 161–168.

Lewis, R.S., Vervoort, J.D., Burmester, R.F., McClelland, W.C, and Chang, Z., 2007, Geochronological constraints on Mesoproterozoic and Neoproterozoic(?) high-grade metasedimentary rocks of north-central Idaho, U.S.A., *in* Link, P.K., and Lewis, R.S., eds., SEPM Special Publication 86, Proterozoic geology of western North America and Siberia, p. 37–53.

Lewis, R.S., Burmester, R.F., Breckenridge, R.M., McFaddan, M.D., and Phillips, W.M., 2008, Preliminary geologic map of the Sandpoint 30 x 60 minute quadrangle, Idaho and Montana, and the Idaho part of the Chewelah 30 x 60 minute quadrangle: Idaho Geological Survey Digital Web Map 94, scale 1:100,000.

Lewis, R.S., Vervoort, J.D., Burmester, R.F., and Oswald, P.J., 2010, Detrital zircon analysis of Mesoproterozoic and Neoproterozoic metasedimentary rocks of north-central Idaho: Implications for development of the Belt-Purcell Basin: Canadian Journal of Earth Sciences, v. 47, no. 11, p. 1383–1404.

Link, P.K., 1998, Regional geologic perspectives on the Belt Supergroup gained since Belt Symposium II, 1983, *in* Berg, R., ed., Proceedings of Belt Symposium III: Montana Bureau of Mines and Geology Special Paper 112, p. 2–10.

Link, P.K., and Fanning, C.M., 2003, Detrital zircon ages from the Yellowjacket, Apple Creek, and Gunsight Formations, Blackbird mining district, Salmon River Mountains, central Idaho: Northwest Geology, v. 32, p. 206–207.

Link, P.K., Winston, D., and Boyack, D., 2003, Stratigraphy of the Mesoproterozoic Belt Supergroup, Salmon River Mountains, Lemhi County, Idaho: Northwest Geology, v. 32, p. 107–133.

Link, P.K., Fanning, C.M., Lund, K.I., and Aleinikoff, J.N., 2007, Detrital zircons, correlation and provenance of Mesoproterozoic Belt Supergroup and correlative strata of east-central Idaho and southwest Montana, *in* Link, P.K., and Lewis, R.S., eds., Proterozoic geology of western North America and Siberia: Society for Sedimentary Geology Special Publication 86, p. 101–128, doi: https://doi.org/10.2110/pec.07.86.0101

Link, P.K., Steele, T., Stewart, E.S., Sherwin, J., Hess, L.R., and McDonald, C., 2013, Detrital zircons in the Mesoproterozoic upper Belt Supergroup in the Beaverhead and Lemhi Ranges, MT and ID: Northwest Geology, v. 42, p. 39–43.

Link, P.K., Stewart, E.D., Steel, T., Sherwin, J., Hess, L.T., and McDonald, C., 2016, Detrital zircons in the Mesoproterozoic upper Belt Supergroup in the Pioneer, Beaverhead, and Lemhi Ranges, Montana and Idaho: The Big White arc, *in* MacLean, J.S., and Sears, J.W., eds., Belt Basin: Window to Mesoproterozoic Earth: Geological Society of America Special Paper 522, Chapter 5, p. 121–138, doi: https://doi.org/10.1130/2016.2522(07)

Lonn, J.D., 2011, Belt rocks and structures of the Lewis and Clark Line in Alberton Gorge, western Montana: A river log: Northwest Geology, v. 40, p. 105–124.

Lonn, J.D., 2014, The northern extent of the Mesoproterozoic Lemhi Group, Idaho and Montana, and stratigraphic and structural relations with Belt Supergroup strata: Geological Society of America Abstracts with Programs, v. 46, no. 5, p. 72.

Lonn, J.D., 2015, Geologic field guide to the Cherry Creek Fault, a possible Proterozoic fault near Helena, Montana: Northwest Geology, v. 44, p. 189–194.

Lonn, J.D., 2017, The Lemhi Group type sections revisited and revised, east-central Idaho: Northwest Geology, v. 46, p. 15–28.

Lonn, J.D., and McFaddan, M.D., 1999, Geologic map of the Montana part of the Wallace 30' x 60' quadrangle: Montana Bureau of Mines and Geology Open File Report 388, scale 1:100,000.

Lonn, J.D., Burmester, R.F., Lewis, R.S., and McFaddan, M.D., 2016a, Giant folds and complex faults in Mesoproterozoic Lemhi strata of the Belt Supergroup, northern Beaverhead Mountains, Montana and Idaho, *in* MacLean, J.S., and Sears, J.W., eds., Belt Basin: Window to Mesoproterozoic Earth: Geological Society of America Special Paper 522, Chapter 6, p. 139–162, doi: https://doi.org/10.1130/2016.2522(06)

Lonn, J.D., Lewis, R.S., Burmester, R.F., and McFaddan, M.D., 2016b, Mesoproterozoic Lemhi strata represent immense alluvial aprons that prograded northwest into the Belt Sea, Idaho and Montana: Geological Society of America Abstracts with Programs, v. 48, no. 6, doi: https://doi.org/10.1130/abs/2016RM-276082

Lonn, J.D., Elliott, C.G., Lewis, R.S., Burmester, R.F., McFaddan, M.D., Stanford, L.R., and Jänecke, S.U., 2019, Geologic map of the Montana part of the Salmon 30' x 60' quadrangle, southwestern Montana: Montana Bureau of Mines and Geology Geologic Map 75, 28 p., 1 sheet, scale 1:100,000.

Luepke, J.J., and Lyons, T.W., 2001, Pre-Rodinian (Mesoproterozoic) super-continental rifting along the western margin of Laurentia: Geochemical evidence from the Belt-Purcell Supergroup: Precambrian Research, v. 111, no. 1, p. 79–90, doi: https://doi.org/10.1016/S0301-9268(01)00157-7

Lydon, J.W., 2000, A synopsis of the current understanding of the geological environment of the Sullivan deposit, *in* Lydon, J.W., Höy, T., Slack, J.F., and Knapp, M.E., eds., The geological environment of the Sullivan deposit, British Columbia: Geological Association of Canada Mineral Deposits Division (MDD) Special Publication 1, p. 12–31.

Lydon, J.W., 2005, Synopsis of the Belt-Purcell basin, https://www.researchgate.net/profile/

John_Lydon2/publication/260302384_SYNOPSIS_OF_THE_BELT-PURCELL_BASIN/links/5446d4990cf2f14fb811cbc9/SYNOPSIS-OF-THE-BELT-PURCELL-BASIN.pdf [Accessed May 4, 2020].

Lydon, J.W., 2007, Geology and metallogeny of the Belt-Purcell Basin, *in* Goodfellow, W.D., ed., Mineral deposits of Canada: A synthesis of major deposit types, district metallogeny, the evolution of geological provinces, and exploration methods: Geological Association of Canada Mineral Deposits Division, Special Publication 5, p. 581–607.

Lydon, J.W., 2010, Tectonic evolution of the Belt-Purcell Basin: Implications for the metallogeny of the Purcell anticlinorium: Geological Survey of Canada Open File 6411, 38 p.

Lydon, J.W., and van Breeman, O., 2013, New zircon ages and their implications for the western margin of the Belt-Purcell Basin of the Purcell Anticlinorium: Northwest Geology, v. 42, p. 45–56.

Lydon, J.W., Höy, T., Slack, J.F., and Knapp, M.E., eds., 2000, The geological environment of the Sullivan deposit, British Columbia: Geological Association of Canada Mineral Deposits Division, Special Publication 1, 834 p.

Lyons, T.W., Frank, T.D., Schreiber, M.E., Winston, D., and Lohmann, K.C., 1998, Geochemical constraints on paleoenvironments with the Belt Supergroup (middle Proterozoic), Montana, *in* Berg, R., ed., Proceedings of Belt Symposium III: Montana Bureau of Mines and Geology Special Paper 112, p. 190–201.

Lyons, T.W., Leupke, J.J., Schreiber, M.E., and Zieg, G.A., 2000, Sulfur geochemical constraints on Mesoproterozoic restricted marine deposition: Lower Belt Supergroup, northwestern United States: Geochemica et Cosmochemica Acta, v. 64, 427–437.

Lyons, T.W., Gallatly, A.M., Frank, T.D., Kah, L.C., Arthur, M.A., and Hurtgen, M.T., 2003, Low sulfate concentrations in the Mesoproterozoic ocean as recorded in carbonates and pyritic shales of the Belt Supergroup and the possible role of methane in the formation of molar-tooth calcite: Northwest Geology, v. 32, p. 209–210.

Mauk, J.F., and White, B.G., 2004, Stratigraphy of the Proterozoic Revett Formation and its control on Ag-Pb-Zn vein mineralization in the Coeur d'Alene district, Idaho: Economic Geology, v. 99, p. 295–312.

McDonald, C., and Lonn, J.D., 2013, Revisions of Mesoproterozoic and Cambrian stratigraphy in the Pioneer and Highland Mountains, southwestern Montana, and resulting implications for the paleogeography of the Belt Basin: Northwest Geology, v. 42, p. 93–102.

McDonald, C., Elliott, C.G., Vuke, S.M., Lonn, J.D., and Berg, R.B., 2012, Geologic map of the Butte South 30′ x 60′ quadrangle, southwestern Montana: Montana Bureau of Mines and Geology Open-File Report 622, scale 1:100,000.

McMannis, W.J., 1963, LaHood Formation—A coarse facies of the Belt Series in southwestern Montana: Geological Society of America Bulletin, v. 74, p. 407–436, doi: https://doi.org/10.1130/0016-7606(1963)74[407:LFCFOT]2.0.CO;2

Miller, F.K., and Yates, R.G., 1976, Geologic map of the west half of the Sandpoint 1° x 2° quadrangle: U.S. Geological Society Open-File Report 76-327, scale 1:125,000.

Mudge, M.R., Earhart, R.L., Whipple, J.W., and Harrison, J.E., 1982, Geologic and structure maps of the Choteau 1° x 2° quadrangle, western Montana: U.S. Geological Survey Miscellaneous Investigations Map I-1300, scale 1:250,000.

Mueller, P., Foster, D., Wooden, J., Mogk, D., and Lewis, R., 2003, Archean and Proterozoic sources for basal quartzites from the eastern and western margins of the Belt Basin: Northwest Geology, v. 32, p. 215–216.

Mueller, P., Mogk, D., Wooden, J., and Spake, D., 2016, U-Pb ages of zircons from the lower Belt Supergroup and proximal crystalline basement: Implications for the early evolution of the Belt Basin, *in* MacLean, J.S., and Sears, J.W., eds., Belt Basin: Window to Mesoproterozoic Earth: Geological Society of America Special Paper 522, p. 283–303, doi: https://doi.org/10.1130/2016.2522(11)

Nesheim, T.O., Vervoort, J.D., McClelland, W.C., Gilotti, J.A., and Lang, H.M., 2012, Mesoproterozoic syntectonic garnet with Belt Supergroup metamorphic tectonites: Evidence of Grenville-age metamorphism and deformation along northwest Laurentia: Lithos, v. 134, p. 91–107, doi: https://doi.org/10.1016/j.lithos.2011.12.008

O'Neill, J.M., 1995, Table Mountain Quartzite and Moose Formation, and associated rocks of the Middle Proterozoic Belt Supergroup, Highland Mountains, Montana: U.S. Geological Survey Bulletin 2121-A, 35 p.

O'Neill, J.M., 1998, Stratigraphic character and structural setting of the Belt Supergroup in the Highland Mountains, southwestern Montana, *in* Berg, R.B., ed., Belt Symposium III: Montana Bureau of Mines and Geology Special Publication 112, p. 12–16.

O'Neill, J.M., Ruppel, E.T., and Lopez, D.A., 2007, Great Divide megashear, Montana, Idaho, and Washington—An intraplate crustal-scale shear zone recurrently active since the Mesoproterozoic: U.S. Geological Survey Open-File Report 2007-1280-A, 10 p.

Overocker, Quintin M., 2006, Origin and classification of Middle Wallace breccias, Mesoproterozoic Belt Supergroup, Montana and Idaho: Knoxville, University of Tennessee, M.S. thesis, 64 p., http://trace.tennessee.edu/utk_gradthes/1759 [Accessed May 6, 2020].

Pearson, R.C., 1996, Cambrian(?), Middle Proterozoic, and Archean rocks penetrated in a borehole near Argenta, Beaverhead County, Montana, and some paleogeographic and structural implications: U.S. Geological Survey Bulletin 2121-B, 15 p., doi: https://doi.org/10.3133/b2121B

Poage, M.A., Hyndman, D.W., and Sears, J.W., 2000, Petrology, geochemistry, and diabase-granophyre relations of a thick basaltic sill emplaced into wet sediments, western Montana: Canadian Journal of Earth Sciences, v. 37, p. 1109–1119, doi: https://doi.org/10.1139/e00-022

Pollock, M.D., Kay, L.C., and Bartley, J.K., 2003, A petrographic test of the gas expansion hypothesis—Implications for the genesis and temporal distribution of molar-tooth structures: Northwest Geology, v. 32, p. 217–218.

Pollock, M.D., Kay, L.C., and Bartley, J.K., 2006, Morphology of molar-tooth structures in Precambrian carbonates: Influence of substrate rheology and implications for genesis: Journal of Sedimentary Research, v. 76, p. 310–323, doi: https://doi.org/10.2110/jsr.2006.021

Pratt, B.R., 1998a, Syneresis cracks: Subaqueous shrinkage in argillaceous sediments caused by earthquake-induced dewatering: Sedimentary Geology, v. 117, p. 1–10.

Pratt, B.R., 1998b, Molar-tooth structure in Proterozoic carbonate rocks: Origin from synsedimentary earthquakes, and implications for the nature and evolution of basins and marine sediment: Geological Society of America Bulletin, v. 110, p. 1028–1045, doi: https://doi.org/10.1130/0016-7606(1998)110%3C1028:MTSIPC%3E2.3.CO;2

Pratt, B.R., 2001, Oceanography, bathymetry, and syndepositional tectonics of a Precambrian intracratonic basin: Integrating sediments, storms, earthquakes and tsunamis in the Belt Supergroup (Helena Formation, ca. 1.45 Ga), western North America: Sedimentary Geology, v. 141–142, p. 371–394.

Pratt, B.R., 2017a, Sedimentary dynamics of the Grinnell Formation (Belt Supergroup, ca. 1.45 Ga), western North America: Not arid playas crossed by creeks but subaqueous muds rocked by earthquakes and swept by tsunamis: Geological Society of America Abstracts with Programs, v. 49, no. 5, doi: https://doi.org/10.1130/abs/2017RM-293144

Pratt, B.R., 2017b, The Mesoproterozoic Belt Supergroup in Glacier and Waterton Lakes National Parks, northwestern Montana and southwestern Alberta: Sedimentary facies and syndepositional deformation, *in* Hsieh, J.C.C., ed., Geologic field trips of the Canadian Rockies: 2017 meeting of the GSA Rocky Mountain Section: Geological Society of America Field Guide 48, p. 123–135, doi: https://doi.org/10.1130/2017.0048(04)

Pratt, B.R., and Ponce, J.J., 2019, Sedimentation, earthquakes, and tsunamis in a shallow, muddy epeiric sea: Grinnell Formation (Belt Supergroup, ca. 1.45 Ga), western North America: Geological Society of America Bulletin, v. 131, no. 9/10, p. 1411–1439, doi: https://doi.org/10.1130/B35012.1

Price, R.A., 2008, Cordilleran tectonics and the evolution of the western Canada Sedimentary Basin, Chapter 2, *in* Mossop, G.D., and Shetsen, I., eds., Geological atlas of the western Canada Sedimentary Basin: Canadian Society of Petroleum Geologists and Alberta Research Council, https://ags.aer.ca/publications/chapter-2-cordilleran-tectonics.html [Accessed May 6, 2020].

Ransome, F.L., and Calkins, F.C., 1908, The geology and ore deposits of the Coeur d'Alene district, Idaho: U.S. Geological Survey Professional Paper 62, 203 p.

Retallack, G.J., Dunn, K.L., and Saxby, J., 2013, Problematic Mesoproterozoic fossil *Horodyskia* from Glacier National Park, Montana, USA: Precambrian Research, v. 226, p. 125–142, doi: https://doi.org/10.1016/j.precamres.2012.12.005

Reynolds, M.W., and Brandt, T.R., 2006a, Preliminary geologic map of the White Sulphur Springs 30' x 60' quadrangle, Montana: U.S. Geological Survey Open-File Report 2006-1329, scale 1:100,000.

Reynolds, M.W., and Brandt, T.R., 2006b, Preliminary geologic map of the Townsend 30' x 60' quadrangle, Montana: U.S. Geological Survey Open-File Report 2006-1138, scale 1:100,000.

Reynolds, M.W., and Brandt, T.R., 2007, Geologic map of the Canyon Ferry Dam 30' × 60' quadrangle, west-central Montana: U.S. Geological Survey Scientific Investigations Map 2860, scale 1:100,000.

Ross, G.M., and Villeneuve, M., 2003, Provenance of the Mesoproterozoic (1.45 Ga) Belt Basin (western North America): Another piece in the pre-Rodinia paleogeographic puzzle: Geological Society of America Bulletin, v. 115, p. 1191–1217, doi: https://doi.org/10.1130/B25209.1

Ross, G.M., Parrish, R.R., and Winston, D., 1992, Provenance and U-Pb geochronology of the Mesoproterozoic Belt Supergroup (northwestern United States): Implications for age of deposition and pre-Panthalassa plate reconstructions: Earth and Planetary Science Letters, v. 113, p. 57–76, doi: https://doi.org/10.1016/0012-821X(92)90211-D

Rule, R., and Pratt, B.R., 2017, *Horodyskia*: Not a fossil but due to sediment-microbial interactions in the Appekunny Formation (Belt Supergroup, ca. 1.46 Ga, western North America): Geological Society of America Abstracts with Programs, v. 49, no. 5, doi: https://doi.org/10.1130/abs/2017RM-293334

Ruppel, E.T., 1993, Cenozoic tectonic evolution of southwest Montana and east-central Idaho: Montana Bureau of Mines and Geology Memoir 65, 62 p.

Ruppel, E.T., and O'Neill, J.M., 2003, Early Tertiary detachment faulting in southwest Montana and east-central Idaho: Northwest Geology, v. 32, p. 222–223.

Ruppel, E.T., O'Neill, J.M., and Lopez, D.A., 1993, Geologic map of the Dillon 1° x 2° quadrangle, Idaho and Montana: U.S. Geological Survey Miscellaneous Investigations Series Map I-1803-H, scale 1:250,000.

Ryan, P.C., and Buckley, S.N., 1998, Sedimentation, strata-bound CuAg mineralization, and syndepositional tectonics in the Revett Formation, Flathead Indian Reservation, western Montana, *in* Berg, R., ed., Proceedings of Belt Symposium III: Montana Bureau of Mines and Geology Special Paper 112, p. 278–289.

Scheiber, J., 1998, Sedimentalogical and mineralogical features of the Belt Supergroup and their bearing on the lacustrine versus marine debate, *in* Berg, R., ed., Proceedings of Belt Symposium III: Montana Bureau of Mines and Geology Special Paper 112, p. 177–189.

Sears, J.W., 1994, Thrust rotation of the Belt Basin, Canada and United States: Northwest Geology, v. 23, p. 81–92.

Sears, J.W., 2000, Rotational kinematics of the Rocky Mountain thrust belt of northern Montana, *in* Schalla, R.A., and Johnson, E.H., eds., Montana/Alberta thrust belt and adjacent foreland: Billings, Montana Geological Society, 50th Anniversary Symposium, v. 1, p. 143–150.

Sears, J.W., 2007a, Belt-Purcell Basin: Keystone of the Rocky Mountain fold-and-thrust belt, United States and Canada, *in* Sears, J.W., Harms, T.A., and Evenchick, C.A., eds., Whence the Mountains? Inquiries into the evolution of orogenic systems: A volume in honor of Raymond A. Price: Geological Society of America Special Paper 433, p. 147–166, doi: https://doi.org/10.1130/2007.2433(07)

Sears, J.W., 2007b, Destabilization of a Proterozoic epicontinental pediment by rifting: A model for the Belt-Purcell Basin, North America, *in* Link, P.K., and Lewis, R.S., eds., Proterozoic geology of western North America and Siberia: SEPM (Society for Sedimentary Geology) Special Publication no. 86, p. 55–64.

Sears, J.W., 2013, Balanced and restored cross-sections of the Belt Basin, Montana and Idaho: Northwest Geology, v. 42, p. 75–78.

Sears, J.W., and Link, P.K., 2007, A field guide to the quartzite of Argenta: Neihart Quartzite equivalent?: Northwest Geology, v. 36, p. 221–230.

Sears, J.W., and Price, R.A., 2000, New look at the Siberian connection: No SWEAT: Geology, v. 28, p. 423-426, doi: https://doi.org/10.1130/0091-7613(2000)28%3C423:NLATSC%3E2.0.CO;2

Sears, J.W., and Price, R.A., 2003, Tightening the Siberian connection to western Laurentia: Geological Society of America Bulletin, v. 115, p. 943–953, doi: https://doi.org/10.1130/B25229.1

Sears, J.W., Chamberlain, K.R., and Buckley, S.N., 1998, Structural and U-Pb geochronologic evidence for 1.47 Ga rifting event in the Belt Basin, western Montana: Canadian Journal of Earth Sciences, v. 35, p. 467–475, doi: https://doi.org/10.1139/cjes-35-4-467

Sears, J.W., Winston, D., and Lonn, J.D., 2003, Belt Symposium IV: Missoula-Ravalli-Perma-St. Regis-Alberton-Missoula, MT: Northwest Geology, v. 32, p. 1–16.

Sears, J.W., Price, R.A., and Khudoley, A.K., 2004, Linking the Mesoproterozoic Belt-Purcell and Udzha Basins across the west Laurentia-Siberia connection: Precambrian Research, v. 129, p. 291–308, doi: https://doi.org/10.1016/j.precamres.2003.10.005

Slotznick, S.P., 2016, Iron mineralology and redox conditions during deposition of the mid-Proterozoic Appekunny Formation, Belt Supergroup, Glacier National Park, *in* MacLean, J.S., and Sears, J.W., eds., Belt Basin: Window to Mesoproterozoic Earth: Geological Society of America Special Paper 522, Chapter 6, p. 221–242, doi: https://doi.org/10.1130/2016.2522(09)

Slotznick, S.P., Zieg, J., Webb, S.M., Kirschvink, J.L., and Fisher, W.W., 2015, Iron mineralogy and redox chemistry of the Mesoproterozoic Newland Formation related to seafloor-hydrothermal sulfide deposits: Northwest Geology, v. 44, p. 55–72.

Smith, A.G., 2016, A review of molar-tooth structures with some speculations on their origin, *in* MacLean, J.S., and Sears, J.W., eds., Belt Basin: Window to Mesoproterozoic Earth: Geological Society of America Special Paper 522, p. 341–363, doi: https://doi.org/10.1130/2016.2522(03)

Stewart, E.D., Link, P.K., Fanning, C.M., Frost, C.D., and McCurry, M., 2010, Paleogeographic implications of non-North American sediment in the Mesoproterozoic upper Belt Supergroup and Lemhi Group, Idaho and Montana, USA: Geology, v. 38, p. 927–930, doi: https://doi.org/10.1130/G31194.1

Stewart, E.D., Steele, E.D., Stewart, D.E., and Link, P.K., 2014, Geologic map of the Gibbonsville, Shewag Lake, and Allan Mountain quadrangles, and parts of the Lost Trail Pass and Big Hole Pass quadrangles, Lemhi County, Idaho, and Ravalli and Beaverhead Counties, Montana: Idaho Geological Survey Technical Report 14-2, scale 1:40,000.

Suttill, H.L., 2009, Sedimentological evolution of the Emine and Kamchia Basins, eastern Bulgaria: Edinburgh, Scotland, University of Edinburgh, M.P. thesis, 239 p., http://hdl.handle.net/1842/4919 [Accessed May 6, 2020].

Thiessen, K.M., and Wallace, C.A., 1996, Origin of carbonate sedimentary breccias of the Wallace Formation, Belt Supergroup, Montana and Idaho: Geological Society of America Abstracts with Programs, v. 28, no. 7, p. 278.

Thorson, J.P., 1993, Gold resources in the Greyson Formation, Big Belt Mountains, Montana, *in* Berg, R.B., compiler, Belt Symposium III Abstracts: Montana Bureau of Mines and Geology Open-File Report 381, p. 56–58.

Tysdal, R.G., 1996a, Geologic map of the Lem Peak quadrangle, Lemhi County, Idaho: U.S. Geological Survey Geologic Quadrangle Map GQ-1777, scale 1:24,000.

Tysdal, R.G., 1996b, Geologic map of parts of the Hayden Creek and Mogg Mountain quadrangles, Lemhi County, Idaho: U.S. Geological Survey Miscellaneous Investigations Series I-2563, scale 1:24,000.

Tysdal, R.G., 2000a, Stratigraphy and depositional environments of middle Proterozoic rocks, northern part of the Lemhi Range, Lemhi County, Idaho: U.S. Geological Survey Professional Paper 1600, 40 p.

Tysdal, R.G., 2002, Structural geology of western part of Lemhi Range, east-central Idaho: U.S. Geological Survey Professional Paper 1659, 33 p.

Tysdal, R.G., 2003, Correlation, sedimentology, and structural setting, upper strata of Mesoproterozoic Apple Creek Formation and lower strata of Gunsight Formation, Lemhi Range to Salmon River Mountains, east-central Idaho: U.S. Geological Survey Professional Paper 1668-A, p. 1–22.

Tysdal, R.G., and Moye, F., 1996, Geologic map of the Allison Creek quadrangle, Lemhi County, Idaho: U.S. Geological Survey Geologic Quadrangle Map GQ-1778, scale 1:24,000.

Tysdal, R.G., Lindsey, D.A., Lund, K.I., and Winkler, G.R., 2005, Alluvial facies, paleocurrents, and source of the Mesoproterozoic Gunsight Formation, east-central Idaho and southwestern Montana: U.S. Geological Survey Professional Paper 1700-B, p. 21–39.

Umpleby, J.B., 1913, Geology and ore deposits of Lemhi County, Idaho: U.S. Geological Survey Bulletin 528, 182 p.

Walcott, C.D., 1899, Precambrian fossiliferous formations: Geological Society of America Bulletin, v. 10, p. 199–244, doi: https://doi.org/10.1130/GSAB-10-199

Wallace, C.A., 1998, Paleotransport directions and basin configurations, middle part of the Missoula Group (Belt Supergroup, middle Proterozoic), western Montana, *in* Berg, R., ed., Proceedings of Belt Symposium III: Montana Bureau of Mines and Geology Special Paper 112, p. 88–103.

Wallace, C.A., Schmidt, R.G., Lidke, D.J., Waters, M.R., Elliott, J.E., French, A.B., Whipple, J.W., Zarske, S.E., Blaskowski, M.J., Heise, B.A., Yeoman, R.A., O'Neill, J.M., Lopez, D.A., Robinson, G.D., and Klepper, M.R., 1986, Preliminary geologic map of the Butte 1° x 2° quadrangle, Montana: U.S. Geological Survey Open File Report 86-292, scale 1:250,000.

Weed, W.H., 1899, Description of the Little Belt Mountains quadrangle (Montana): U.S. Geological Survey Geological Atlas, Folio 56, 9 p.

Whipple, J.W., and Morrison, G.G., 1993, Stratabound gold in the Greyson Formation, Big Belt Mountains, Montana, *in* Berg, R.B., compiler, Belt Symposium III Abstracts: Montana Bureau of Mines and Geology Open-File Report 381, p. 65–66.

White, B.G., 2000, Coeur d'Alene mining district: Product of preconcentrated source deposits and tectonism within the Lewis and Clark Line, *in* Roberts, S., and Winston D., eds., Geologic field trips, western Montana and adjacent areas: Rocky Mountain Section of the Geological Society of America, University of Montana, p. 95–101.

White, B.G., and Appelgate, L., 1999, Syndepositional faulting in the western Belt Basin as localizer of preconcentrated metal sources for Coeur d'Alene silver-lead deposits: Geological Society of America Abstracts with Programs, v. 31, no. 4, p. A-61.

White, B.G., and Appelgate, L., 2000, Localization of the Coeur d'Alene Mining District at the juncture of the Lewis and Clark and Noxon Lines: Geological Society of America Abstracts with Programs, v. 32, no. 5, p. A-42.

White, B.G., Browne, J., and Appelgate, L., 2000, The Noxon Line: A longitudinal arch in the central Belt Basin: Geological Society of America Abstracts with Programs, v. 32, no. 5, p. A-42.

Willis, B., 1902, Stratigraphy and structure, Lewis and Livingston Ranges, Montana: Geological Society of America Bulletin, v. 13, p. 305–352, doi: https://doi.org/10.1130/GSAB-13-305

Winston, D., 1986a, Sedimentation and tectonics of the Middle Proterozoic Belt Basin and their influence on Phanerozoic compression and extension in western Montana and northern Idaho, *in* Peterson, J.A., ed., Sedimentation and tectonics: American Association of Petroleum Geologists Memoir 41, p. 87–118.

Winston, D., 1986b, Stratigraphic correlation and nomenclature of the Middle Proterozoic Belt Supergroup, Montana, Idaho, and Washington, *in* Roberts, S.M., ed., Belt Supergroup: Montana Bureau Mines and Geology Special Paper 94, p. 69–84.

Winston, D., 1986c, Sedimentology of the Ravalli Group, Middle Belt carbonate, and Missoula Group, Middle Proterozoic Belt Supergroup, Montana, Idaho, and Washington, *in* Roberts, S.M., ed., Belt Supergroup: Montana Bureau Mines and Geology Special Paper 94, p. 85–124.

Winston, D., 1998, Lacustrine cycles in the Helena and Wallace Formations, Middle Proterozoic Belt Supergroup, *in* Berg, R., ed., Proceedings of Belt Symposium III: Montana Bureau of Mines and Geology Special Paper 112, p. 70–86.

Winston, D., 2003, Proposed revision of the Helena and Wallace Formations, mid-Proterozoic Belt Supergroup: Northwest Geology, v. 32, 172–185.

Winston, D., 2007, Revised stratigraphy and depositional history of the Helena and Wallace Formations, Mid-Proterozoic Piegan Group, Belt Supergroup, Montana and Idaho, *in* Link, P.K., and Lewis, R.S., eds., Proterozoic geology of western North America and Siberia: SEPM (Society for Sedimentary Geology) Special Publication 86, p. 65–100.

Winston, D., 2013, Analysis of sheetflood fluvial systems in the Middle Proterozoic Belt Supergroup, northern Idaho and northwestern Montana: Northwest Geology, v. 42, p. 71–74.

Winston, D., 2016, Sheetflood sedimentology of the Mesoproterozoic Revett Formation, Belt Supergroup, northwestern Montana, USA, *in* MacLean, J.S., and Sears, J.W., eds., Belt Basin: Window to Mesoproterozoic Earth: Geological Society of America Special Paper 522, p. 341–363, doi: https://doi.org/10.1130/2016.2522(01)

Winston, D., and Link, P.K., 1993, Middle Proterozoic rocks of Montana, Idaho and eastern Washington: The Belt Supergroup, *in* Reed, J.C., Jr., Bickford, M.E., Houston, R.S., Link, P.K., Rankin, D.W., Sims, P.K., and Van Schmus, W.R., eds., Precambrian: Conterminous U.S.: Geological Society of America, Decade of North American Geology, v. C-2, p. 487–517.

Winston, D., and Sears, J.W., 2013, Stratigraphy of the Proterozoic Belt Supergroup and structure of the Belt Basin: Glacier National Park and Blackfoot River Canyon, Montana: Northwest Geology, v. 42, p. 237–278.

Winston, D., and Smith, S.V., 2016, Crinkle cracks in the Proterozoic Piegan Group, Belt Supergroup, Montana and Idaho: A descriptive style of sand-filled cracks hypothetically formed by subaqueous solitary-like waves, *in* MacLean, J.S., and Sears, J.W., eds., Belt Basin: Window to Mesoproterozoic Earth: Geological Society of America Special Paper 522, p. 341–363, doi: https://doi.org/10.1130/2016.2522(02)

Winston, D., and Wheeler, T., 2006, Shepard, Mount Shields, and Bonner Formations of the Middle Proterozoic Missoula Group in Libby syncline, Kootenai Falls, northwest Montana: Northwest Geology, v. 35, p. 63–76.

Winston, D., Link, P.K., and Hathaway, N., 1999, The Yellowjacket is not the Prichard and other heresies: Belt Supergroup correlations, structure and paleogeography, east-central Idaho, *in* Hughes, S.S., and Thackray, G.D., eds., Guidebook to the geology of eastern Idaho: Pocatello, Idaho Museum of Natural History, p. 3–20.

Winston D., Rittel, J.F., and Furniss, G., 1999, Gas bubble and expansion crack origin of molar-tooth calcite structures in the Middle Proterozoic Belt Supergroup, western Montana: Discussion and reply: Journal of Sedimentary Research, v. 69, p. 1141–1145, doi: https://doi.org/10.2110/jsr69.1136

Zieg, J., Scartozzi, V., Chutas, N., Albers, D., Gostomski, K., and Jones, J., 2013, Black Butte copper deposits, lower Belt Supergroup, Montana: Northwest Geology, v. 42, p. 131-148.

Zirakparvar, N.A., Vervoort, J.D., McClelland, W., and Lewis, R.S., 2010, Insights into the metamorphic evolution of the Belt-Purcell Basin: Evidence from Lu-Hf garnet geochronology: Canadian Journal of Earth Sciences, v. 47, p. 161–179, doi: https://doi.org/10.1139/E10-001