Regional structure and kinematic history of the Cordilleran fold-thrust belt in northwestern Montana, USA

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ABSTRACT

The Cordilleran thrust belt of northwestern Montana (United States) has received much less attention than its counterparts in the western interior of USA and Canada. The structure of the thrust belt in this region is well preserved and has not been strongly overprinted by Cenozoic extension, providing an opportunity to reconstruct its geometry and to relate it to the foreland basin system. The thrust belt in this region consists of a frontal part of highly deformed Paleozoic, Mesozoic, and Paleocene sedimentary rocks, and a western region dominated by a >15-km-thick succession of Proterozoic Belt Supergroup strata underlain by faults of the Lewis thrust system. The frontal part can be subdivided into the foothills and the Sawtooth Range. At the surface, the foothills show deformed Mesozoic and Paleocene rocks; at depth, reflection seismic data indicate numerous thrust faults carrying Paleozoic strata. The Sawtooth Range, south from the Lewis thrust salient, is defined by steeply dipping imbricate thrusts that detach at the basal Cambrian stratigraphic level. The Sawtooth Range plunges northward beneath the Lewis thrust salient and diverges into a pair of independent thrust systems that form the Flathead and Waterton duplexes in Canada. The relatively minor internal deformation in the western part of the thrust belt resulted from the great rheological strength of the Belt Supergroup rocks and initial high taper of the preorogenic stratigraphic wedge. A new ~145-km-long balanced cross section indicates ~135 km of shortening, a value similar to that in the southern part of the Canadian thrust belt. Previous work and new conventional and isotopic provenance data from the foreland basin and U-Pb ages from crosscutting intrusive rocks establish a preliminary kinematic model for this segment of the Cordilleran thrust belt. The emerging pattern is a relatively simple forelandward progression of thrusting events. Most shortening in the Lewis thrust system, Sawtooth Range, and foothills occurred roughly between mid-Campanian and Early Eocene time (ca. 75–52 Ma), yielding a shortening rate of ~5.9 mm/yr. This pattern differs from the pattern of shortening in the better known Sevier thrust belt to the south, where regional far-traveled Proterozoic quartzite-bearing thrust sheets were mainly active during Early Cretaceous time. From Middle Eocene to Early Miocene time, this sector of the Cordillera collapsed, generating a number of extensional depocenters.

INTRODUCTION

Classic studies of the Cordilleran orogenic belt in the USA and Canada were among the first to demonstrate modern concepts of thrust belt structure, kinematics, and foreland basin development (e.g., Bally et al., 1966; Armstrong, 1968; Dahlstrom, 1970; Price and Mountjoy, 1970; Price, 1973, 1981; Gordy et al., 1977; Royse et al., 1975; Burchfiel and Davis, 1972, 1975; Jordan, 1981; Lamerson, 1982; Wiltschko and Dorr, 1983). However, these and subsequent studies were concentrated in two large regions: the Sevier thrust belt of southern Nevada, Utah, and Wyoming, USA (e.g., Burchfiel and Davis, 1972; Royse et al., 1975; Lamerson, 1982; Wiltschko and Dorr, 1983; Coogan, 1992; Lawton, 1994; DeCelles, 1994; Mitra, 1997; Constenius et al., 2003; DeCelles and Coogan, 2006; Schelling et al., 2007; Yonkee and Weil, 2010), and the frontal Canadian thrust belt in southern Alberta and British Columbia, Canada (e.g., Bally et al., 1966; Gordy et al., 1977; Price, 1981, 1994, 2000; Fermor and Moffat, 1992; McMechan and Thompson, 1993; Fermor, 1999). This focused work left a large region in northwestern Montana relatively undocumented in terms of structure, kinematic history, and its relationship with the greater Cordilleran thrust belt (DeCelles, 2004). This region contains significant structural features that have no apparent counterparts in other parts of the Cordilleran thrust belt. Of particular interest is the Lewis thrust system, an immense thrust system that has tectonically transported a slab of rock >7 km in thickness more than 100 km toward the foreland region. In contrast to the other major Precambrian rock–carrying thrust faults in the Cordillera, such as the Paris, Willard, Sheep Rock, and Canyon Range thrusts, which formed in the interior part of the thrust belt in Idaho and Utah during the Early Cretaceous, the Lewis is a frontal thrust system that sustained motions as late as Late Paleocene–Early Eocene. However, surprisingly little was previously known regarding the evolution of this thrust system as recorded in the foreland basin stratigraphy and unraveled through detailed cross-section balancing.

During the 1960s and 1970s, the frontal part of the northwestern Montana thrust belt was mapped by M.R. Mudge and collaborators (see Mudge et al., 1982; Mudge and Earhart, 1983) under the auspices of the U.S. Geological Survey (USGS), resulting in a set of 1:24,000 quadrangle maps and large-scale compilations. Subsequently, 1° × 2° quadrangle maps of the northernmost part of this segment of the thrust belt and the hinterland regions were compiled by Harrison et al. (1986, 1992, 1998), based on previous work. The region was mostly mapped just prior to the widespread acceptance of modern ideas about thrust belt systematics and timing (Bally et al., 1966; Dahlstrom, 1970; Boyer and Elliott, 1982) and the relationships between thrust belts and foreland basin systems (Jordan, 1981; Beaumont, 1981). As a result, the early structural cross sections are not balanced (e.g., Harrison et al., 1980; Mudge, 1982; Mudge et al., 1982; Mudge and Earhart, 1983) and virtually no attempt has been made to integrate the

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structural geometry and kinematics with the subsidence history and provenance of the adjacent foreland basin. Subsequent published structural cross sections have been very local (e.g., Mitra, 1986; Holl and Anastasio, 1992), of very low resolution in the frontal part of the thrust belt (e.g., Fritts and Klipping, 1987; Sears, 2001, 2007), or unbalanced (e.g., Harrison et al., 1992).

Our primary objectives are to document the regional-scale structure and kinematic history of the northwestern Montana thrust belt, and to link it with the evolution of the foreland basin. An almost complete record of sedimentation in the foreland basin system, local igneous rocks and crosscutting relationships, and excellent subsurface data provide control on the major kinematic events in the thrust belt. We find that major structural elements in northwestern Montana were controlled by preexisting stratigraphic architecture and lithology, and that the kinematic history of this region is significantly different from that farther south in the Sevier belt.

REGIONAL SETTING

The Cordilleran orogenic belt of western North America formed during Jurassic–Eocene time in response to convergence between Pacific domain plates and the North American plate (Monger et al., 1982; Burchfiel et al., 1992; Saleeby, 1992; Allmendinger, 1992; Coney and Evenchick, 1994; DeCelles, 2004; Dickinson, 2004). In northwestern USA and Canada, subduction was accompanied by accretion of fringing arcs and parautochthonous and exotic terranes (Coney et al., 1980; Monger et al., 1982; Coney and Evenchick, 1994; Dickinson, 2004; Evenchick et al., 2007; Colpron et al., 2007; Ricketts, 2008). The terrane structure and evolution are better understood in southwestern Canada than in the USA Pacific Northwest. These terranes disappear southward beneath the Columbia River Basalt Group (Fig. 1), but correlative terranes are present in the Blue Mountains of northeastern Oregon (Dorsey and Lamaskin, 2007). The current view is that the Pacific Northwest of the USA and adjacent Canada share a similar early accretion history (Dickinson, 2004). In Canada, the easternmost part of these terranes is grouped into a composite unit, the Intermontane superterrane, which was accreted to the North American plate by the Middle to Late Jurassic (Dickinson, 2004; Colpron et al., 2007). The Insular superterrane, the westernmost element in the Cordilleran collage at these latitudes, had a long and complex accretion history that started during the Middle Jurassic (Colpron et al., 2007; Ricketts, 2008).

Under this regime of subduction and collision, a fold-thrust belt developed in the retroarc region; in northwestern Montana (Fig. 2), it involved rocks of three major tectonostratigraphic packages on top of autochthonous North American cratonic basement (Fig. 3). The Belt Supergroup comprises a thick (>15 km) Proterozoic succession of clastic, carbonate, and igneous rocks (Harrison, 1972; Harrison et al., 1974), possibly deposited in an intracontinental rift (Cressman, 1989; Price and Sears, 2000), or a backarc extensional or strike-slip basin (Ross and Villeneuve, 2003). On top of these deposits, Paleozoic strata consist of shelfal carbonate and relatively minor clastic rocks, deposited after the Precambrian rifting event (McMannis, 1965; Bond et al., 1985; Poole et al., 1992). Jurassic–Paleocene mainly clastic sedimentary rocks recycled from rocks of the growing Cordilleran orogenic belt were deposited in a foreland basin to the east; these strata eventually became incorporated into thrust sheets in the frontal part of the thrust belt.

Western regions of the retroarc thrust belt are dominated by rocks of the Belt Supergroup (Figs. 1 and 2). The leading edge of the structures involving Belt rocks is defined by the Lewis thrust, and the related Eldorado and Steinbach thrusts to the south (Fig. 2). The frontal part of the thrust belt is characterized by strongly deformed Paleozoic and Mesozoic strata of the Sawtooth Range and foothills structures. The Sawtooth Range is composed of thrust sheets detached at Cambrian to Mississippian levels. The foothills region is characterized by deformed Mesozoic strata, with blind structures involving Paleozoic rocks at depth. The foothills, Sawtooth Range, and the frontal structures involving Belt Supergroup strata are commonly referred to as the Montana Disturbed Belt (Mudge, 1982).

At the latitude of Helena, the thrust belt forms a pronounced salient that has been interpreted as the result of the inversion of an eastward prolongation of the Belt basin, perhaps a failed rift (Harrison et al., 1974; Winston, 1986; Price and Sears, 2000; Sears and Hendrix, 2004). A complex array of strike-slip faults, known as the Lewis and Clark line, bounds the Helena salient to the north, and extends westward into the hinterland (Figs. 1 and 2). The Lewis and Clark line has a long history of recurrent movement since Precambrian time (Wallace et al., 1990). McClelland and Oldow (2004) interpreted the Lewis and Clark line as an oblique ramp connecting basement-detached hinterland structures of the Shuswap complex in southern British Columbia and Washington with the frontal Laramide uplifts in the USA.

The Middle Jurassic to Early Eocene foreland basin system (Fig. 4) preserves a thickness of strata of as much as 2.5–3 km, but a considerable thickness of Paleocene–Early Eocene proximal deposits was erosionally removed during the Cenozoic (Hardebol et al., 2009). Jurassic strata are characteristically thin and tabular, and possibly reflect deposition in a distal, backbulge depozone (Fuentes et al., 2009). These deposits are bounded above by a regional-scale unconformity that may be partly related to the migration of the flexural forebulge in response to shortening and propagation of the thrust belt into the foreland (Fig. 4). The bulk of the preserved foreland basin fill was deposited in a foredeep depozone. Strata of the wedge-top depozone are not preserved along the thrust belt front, but potentially correlative distal deposits represented by the Fort Union and Wasatch Formations are preserved in isolated localities on the Great Plains. Starting in Eocene time, the hinterland region was overprinted by extensional collapse and subsequently by Basin and Range extension (Constenius, 1996).

THRUST BELT STRUCTURE

The classic southern Canadian tectonomorphic zones of the thrust belt, characterized by distinct topography, stratigraphy, and structural style, continue southward into Montana with some minor differences. The Foothills compose the frontal part of the thrust belt and mainly consist of strongly deformed, imbricated thrust slices of Cretaceous–lower Cenozoic strata that detach from Paleozoic deposits at depth, in a manner similar to the Alberta Foothills (Ross, 1959; Mudge, 1982). The conspicuous frontal syncline along the triangle zone in Canada, i.e., the Alberta syncline (Price, 1981; Jones, 1982), is less well represented in Montana.

The Sawtooth Range is partially equivalent to the Front Ranges of southern Canada (Price, 1981). It is composed of closely spaced imbricated panels of Paleozoic and Mesozoic rocks (McMannis, 1965; Mudge, 1982). The regional detachment of the Sawtooth Range structures cuts downsection progressively to the west, eventually reaching the Cambrian level (Mudge and Earhart, 1983; Boyer, 1992; Harrison et al., 1998). Whereas the Sawtooth Range involves exclusively Phanerozoic rocks, in Canada frontal thrusts involving Belt rocks are included within the Front Ranges.

The dominant structure in northwestern Montana is the Lewis thrust and the related Eldorado and Steinbach thrusts (Fig. 2). The hanging wall of this thrust system consists of several kilometers of Proterozoic and Lower Paleozoic rocks, which are in thrust contact with complexly deformed footwall Paleozoic and Mesozoic strata of the Sawtooth Range (Mudge, 1982; Sears, 2001).

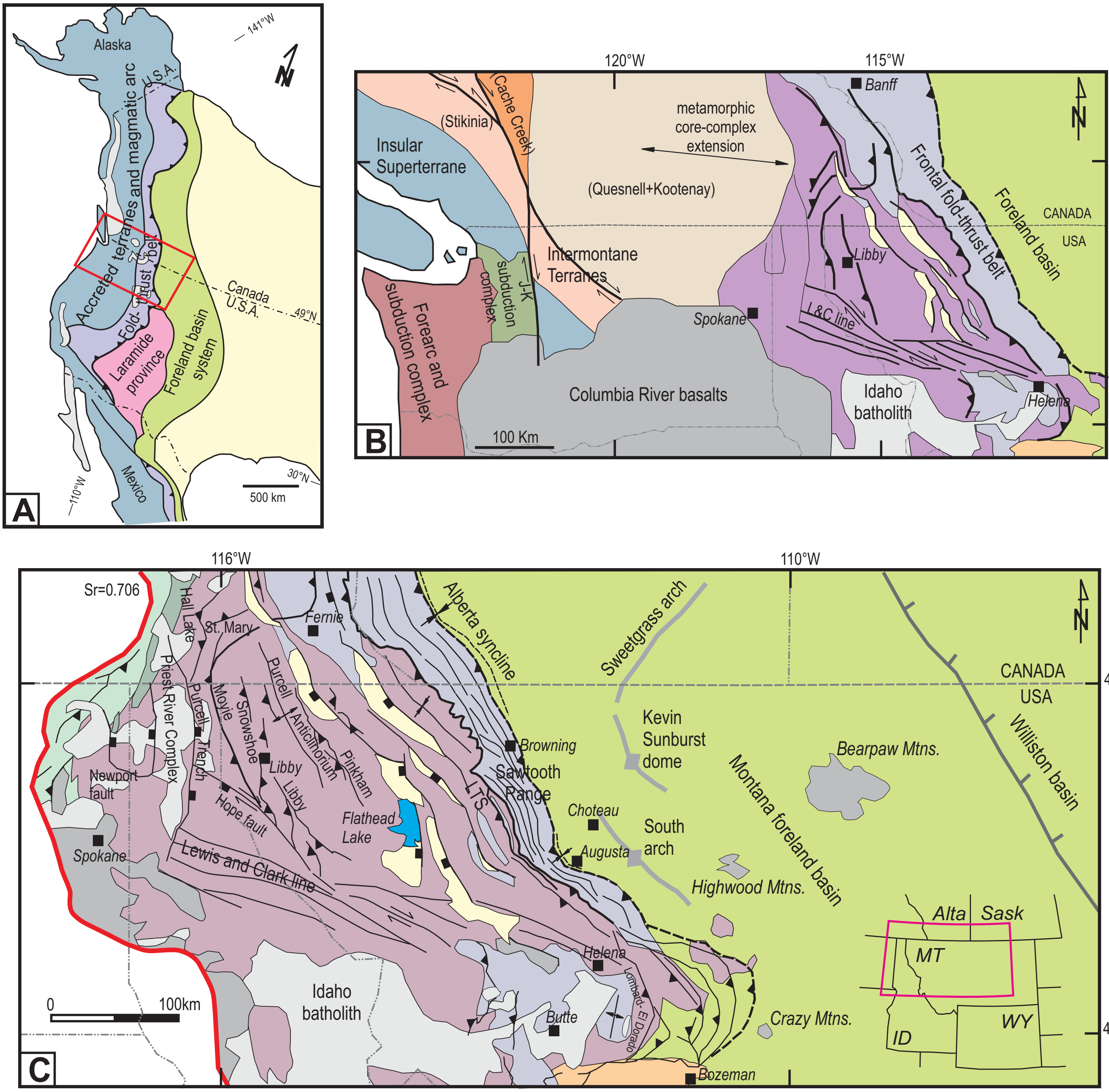
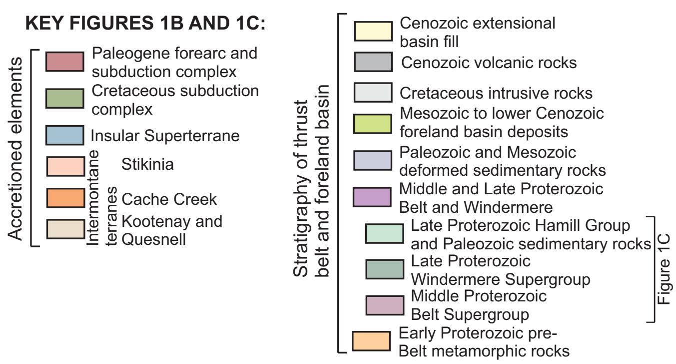


Figure 1. (A) Simplified tectonic map of the North America Cordilleran orogenic system. Box shows location of map of Figure 1B. (B) Terrane map of northwest USA and southwest Canada (terranes based on Dickinson, 2004; Colpron et al., 2007). J-K—Jurassic–Cretaceous; L&C—Lewis and Clark line. (C) Tectonic-index map of the fold-thrust belt and foreland basin of northwestern Montana (MT) and adjacent areas (ID—Idaho; WY—Wyoming). Only names of major thrusts and thrust systems are indicated. LTS—Lewis thrust system (Lewis-Eldorado-Steinbach thrusts). Thrusts in the Sawtooth Range and foothills regions of Montana and Canada are schematic and not ornamented. Compiled from Alpha (1955), Mudge et al. (1982), Harrison et al. (1986, 1992, 1998), Wheeler and McFeely (1991), Constenius (1996), Doughty and Price (2000), Lageson et al. (2001) and Vuke et al. (2007). Figures modified from Fuentes et al. (2011).



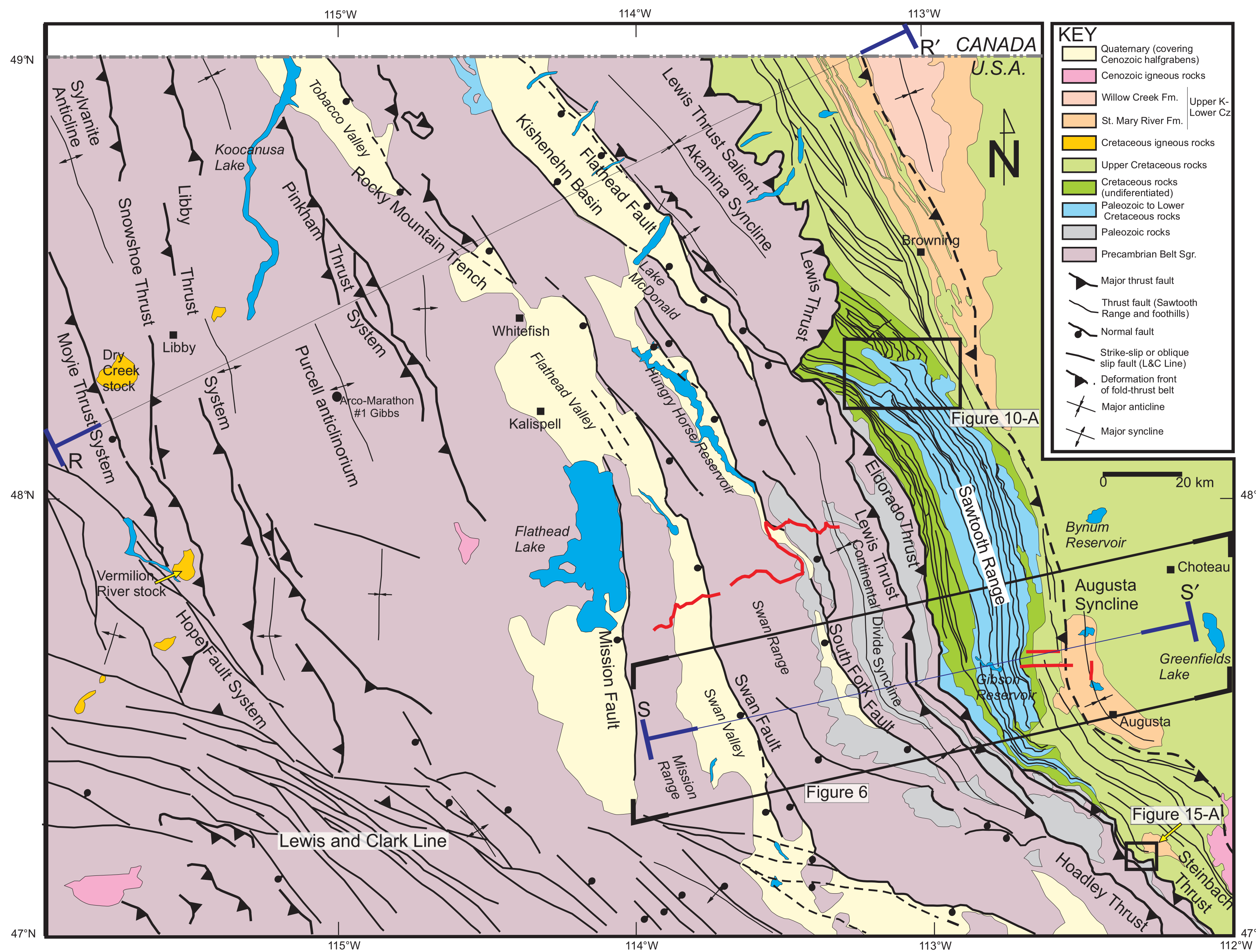
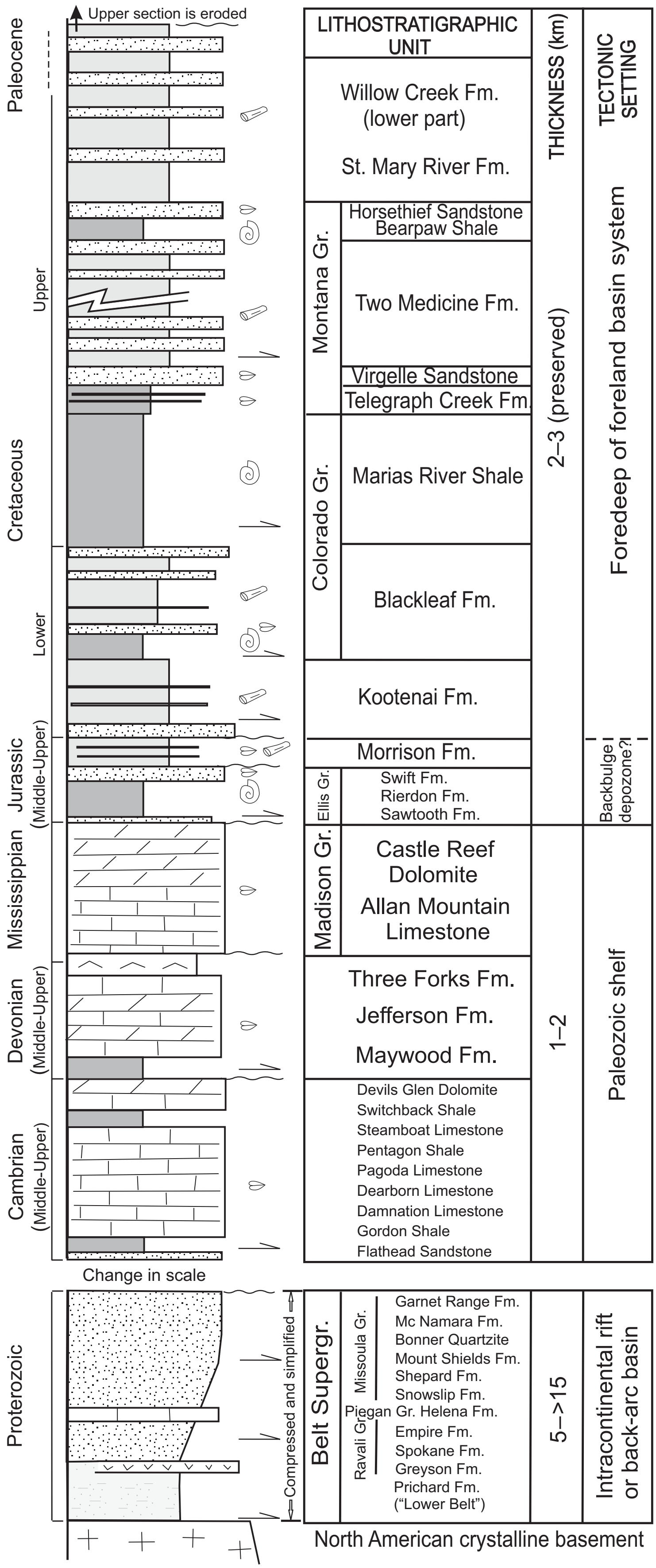


Figure 2. Simplified geologic map of northwest Montana compiled from U.S. Geological Survey 1° × 2° quadrangle maps (Mudge et al., 1982; Harrison et al., 1986, 1992, 1998) and the geologic map of Montana (Vuke et al., 2007). Only major faults of the Moyie, Horseshoe, Libby, and Pinkham thrust systems are displayed. Box in the southeast quadrant of the map indicates location of map of Figure 6. Box in the northern part of the Sawtooth Range show location of map of Figure 10A. Box in southeast corner indicates location of Figure 15A. Red lines indicate seismic lines discussed in the text. Locations of cross sections discussed in the text are also indicated. Sgr—supergroup; L&C—Lewis and Clark.

Figure 3. Simplified stratigraphy of the thrust belt frontal part in northwestern Montana. Lithostratigraphic units after Mudge (1972). Main detachment horizons are indicated by arrows. Note change in scale for Belt Supergroup. Logs: continental; bivalves: shallow marine; ammonites: deep marine deposits.



In this paper, we refer to the frontal thrusts carrying Belt Supergroup rocks as the Lewis thrust system; when referring to individual thrusts (Lewis, Eldorado, and Steinbach thrusts), we use their names as they were mapped by USGS geologists (e.g., Mudge and Earhart, 1980).

To the west, Belt Supergroup rocks are deformed by broad open folds and widely spaced thrusts, in contrast to the foothills and the Sawtooth Range. This region contains a series of thrust systems with poorly known geometry because of postorogenic extensional faulting and Cenozoic sedimentary cover. Thrust systems of this region include the Pinkham, Libby, Snowshoe, and Moyie (Fig. 2). Mid-Cenozoic extensional collapse of the orogenic belt and subsequent Basin and Range deformation produced a number of half-grabens in the interior region of the thrust belt (Fig. 2).

Figure 5 shows a simplified crustal-scale cross section at the latitude of the Lewis thrust salient, south of the international border (Constenius, 1996). The internal structure of the thrust belt in Paleozoic and Mesozoic rocks in the frontal part of the section is not displayed because of scale limitations. As shown in this figure and mentioned above, the dominant feature of the thrust belt is the thrust sheet of Belt Supergroup rocks in the Lewis thrust system hanging wall. The giant Purcell anticlinorium possibly represents a fault-bend fold developed over a major ramp separating autochthonous from allochthonous Belt Supergroup rocks. Western thrust systems have relatively minor slip, and were severely disrupted by Cenozoic extension. The thrust belt basal detachment was reactivated during Cenozoic extension, and listric normal faults fed slip down to it (Constenius, 1996). Regional detachment faulting and metamorphic core complex development to the west produced basement upwarping in the hinterland (Fillipone and Yin, 1994; Doughty and Price, 2000; Johnson, 2006).

Balanced Cross Section

We constructed a new regional balanced cross section to illustrate the geometry of the thrust belt in northwestern Montana. The section is

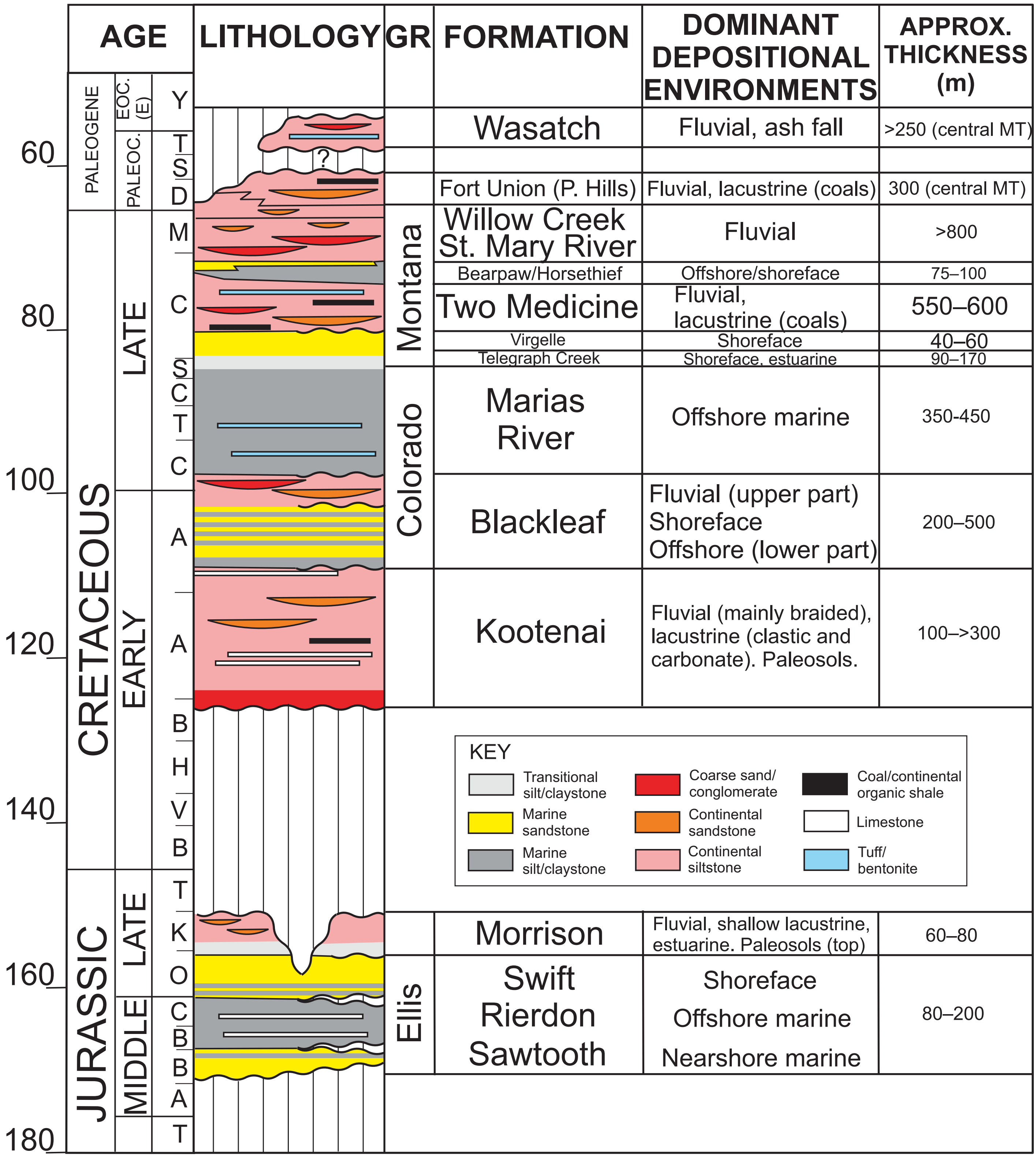


Figure 4. Simplified Jurassic–Early Eocene (EOC.) stratigraphy of northwestern Montana (MT) (modified from Fuentes et al., 2011). P. Hills—Porcupine Hills.

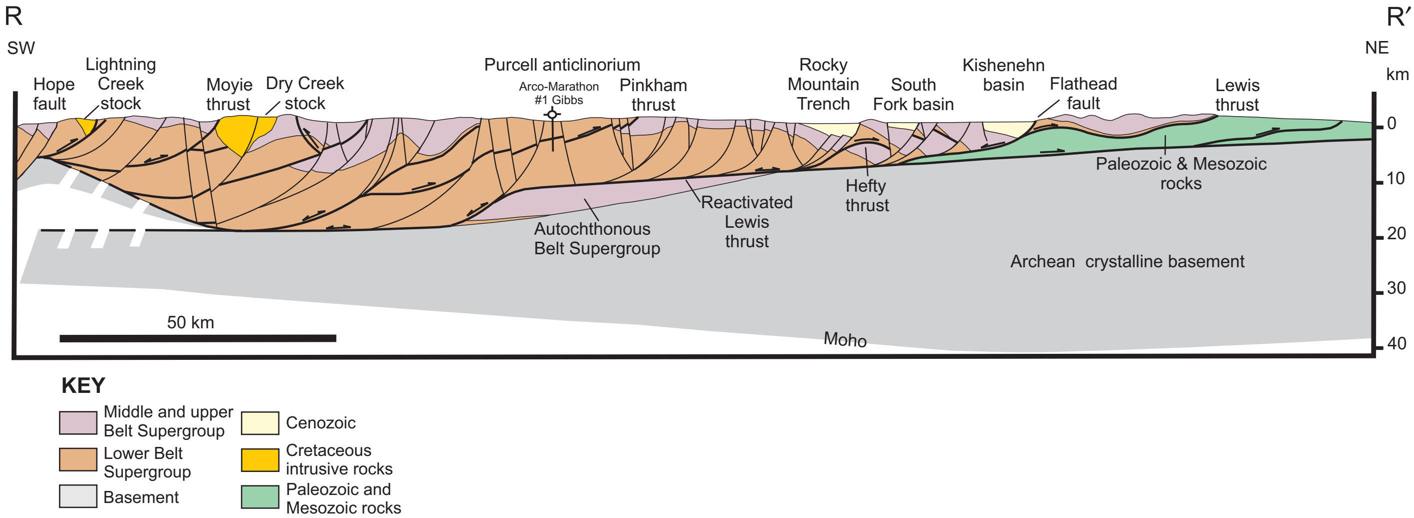


Figure 5. Simplified regional geologic cross section R–R′ of northwest Montana based on results of seismic reflection and refraction profiles, Bouguer gravity data, well control, and balanced cross sections (Bally et al., 1966; Harris, 1985; Fritts and Klipping, 1987; Constenius, 1981, 1988; Boberg et al., 1989; Yoos et al., 1991; Harrison et al., 1992; Sears and Buckley, 1993; Van der Velden and Cook, 1994) (based on Constenius, 1996). The Dry Creek stock is projected southward to the cross section to show its relation with the Moyie thrust.

located in the south-central part of the Sawtooth Range, and extends from the undeformed foreland east of the Augusta syncline westward to the Mission Range, where structures of the thrust belt interfere with structures related to the Lewis and Clark strike-slip system (Figs. 2 and 6). The location of the cross section was chosen in part because no regional, detailed, balanced cross sections exist for the area. Existing sections are local (e.g., Mitra, 1986; Holl and Anastasio, 1992), unbalanced (e.g., Mudge et al., 1982; Mudge and Earhart, 1983), or have poor resolution in the structures east of the Lewis thrust system (e.g., Fritts and Klipping, 1987; Sears, 2001, 2007). Previous work portraying the frontal part of the thrust belt lacked subsurface control (e.g., Sears et al., 2005). Moreover, published USGS geological maps of the Sawtooth Range and the foothills in this region are superb (Mudge, 1965, 1966a, 1966b, 1968), and we have obtained high-quality industry seismic lines and well data from this area.

Methods

The cross section is based on published USGS geological maps and our own mapping. A number of 1:24,000 scale quadrangle maps and a subsequent compilation (Mudge, 1965, 1966a, 1966b, 1968, 1972) provide data from the frontal syncline up to the leading edge of the Lewis thrust system. Maps at 1:125,000 and 1:250,000 scales (Mudge et al., 1982; Mudge and Earhart, 1983) were used in the mostly undeformed plains, and along most of the Lewis thrust sheet, where the structural wavelengths are large. The cross section is oriented ~N80°E, perpendicular to the regional strike direction and approximately parallel to the direction of shortening of the major thrust faults (Fig. 6). The cross section is line-length balanced.

Industry reflection seismic profiles and oil and gas wells were used to interpret the foothills subsurface structure. Most of these seismic profiles were shot during the early 1980s; we reprocessed a 10 km section of a key dipline (Fig. 6). Formation tops from the Montana Power Co. State 1 well (Section 15, Township 21N, Range 7W, Lewis and Clark County) (Fig. 6) were used to tie the seismic data via a synthetic seismogram. Correlation information for the deepest stratigraphic interval was obtained using a synthetic seismogram from the Oxy Larson 1 (Section 1, Township 22N, Range 5W, Teton County), a well to the east that reached basement. A synthetic seismogram from the Husky Gulf Mercier 1 well to the north (Section 32, Township 32N, Range 11W, Glacier County) provided additional control in the identification of stratigraphic horizons from the seismic data. Two contiguous seismic profiles across the Lewis thrust sheet in the Swan valley (Fig. 2) facilitated geometric reconstruction of the hinterland part of the section and in the determination of depth to undeformed basement. The reasoning, justification, and uncertainties behind different aspects of the cross section are discussed in the following and explained in Figure 7.

Results

Figure 7 shows an ~145-km-long balanced cross section of the northwest Montana thrust belt. Petroleum wells drilled in the foothills and seismic data in frontal and western areas of the thrust belt indicate that the base of the Cambrian section dips ~3.3° beneath the foothills and Sawtooth Range, and ~4° beneath the Lewis thrust sheet. A gradual westward increase in basement dip was noted both in Canada (Bally et al., 1966; Dahlstrom, 1970; Fermor, 1999) and the USA (Boyer, 1995).

Seismic reflection profiles from the foothills region show a stacking of thrust sheets containing mechanically strong Paleozoic strata (Fig. 8; Supplemental Figure 1[1]). Based on calibration with synthetic seismograms from nearby wells, the basal detachment for these structures is inferred near the base of the Devonian section. An additional indicator comes from published USGS geological maps (Mudge, 1965, 1968) showing that the Home thrust is located directly below Devonian Jefferson Formation strata. Upper detachment levels for these structures are controlled by fine-grained clastic deposits of the Cretaceous Kootenai Formation, Blackleaf Formation, and the Marias River Shale. The western flank of the frontal synclinorium, referred to as the Augusta syncline, seems to be the result of the emplacement of thrust sheets that, at shallow levels, juxtapose strata of the Marias River Shale on top of the lower Two Medicine Formation. Long flat-on-flat relationships between

[1]Supplemental Figure 1. Color version of seismic section displayed in Figure 8. Vertical scale is two-way traveltime. Location is in Figure 6 (SL 1-a). If you are viewing the PDF of this paper or reading it offline, please visit http://dx.doi.org/10.1130/GES00773.S1 or the full-text article on www.gsapubs.org to view Supplemental Figure 1.

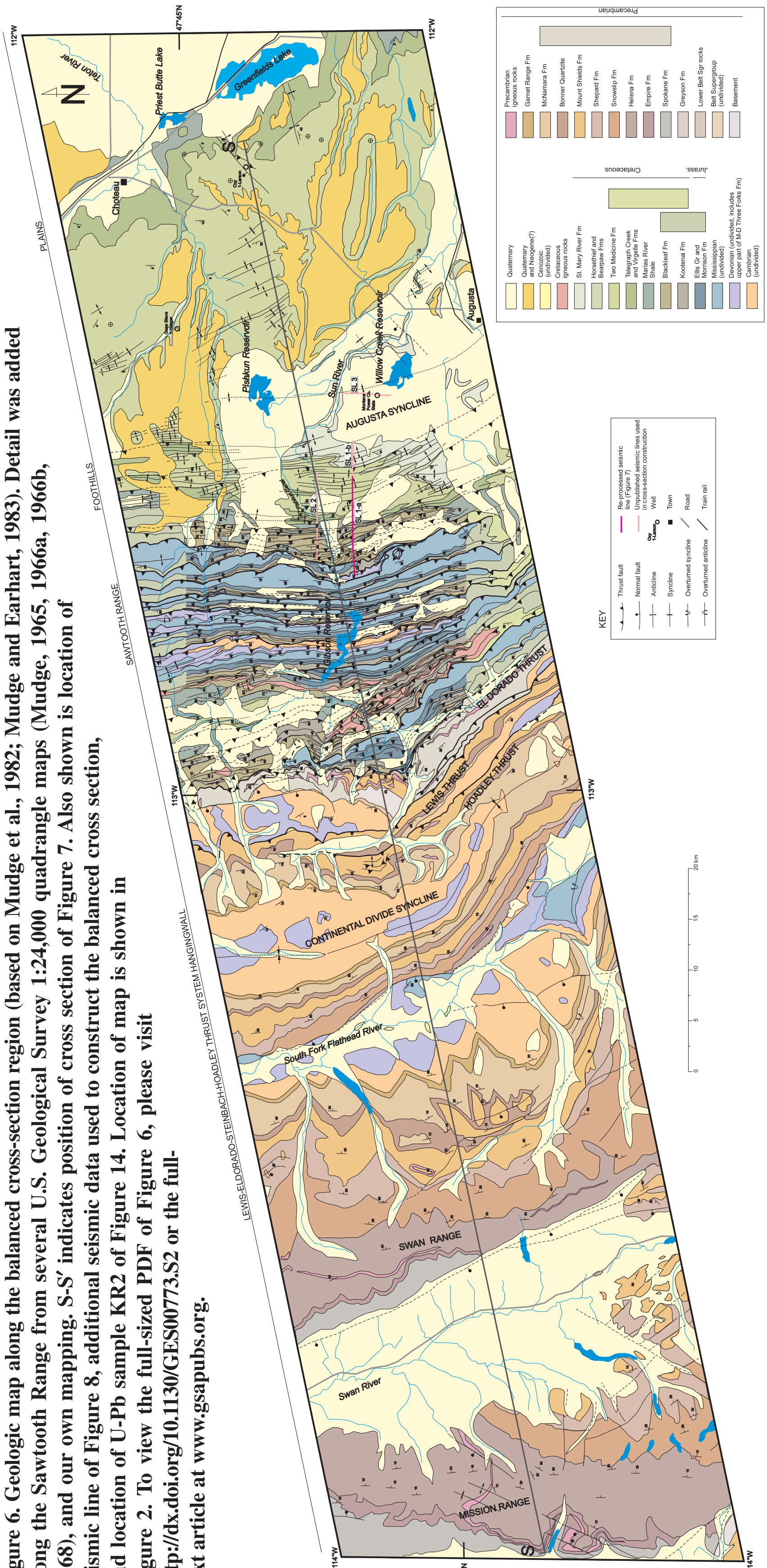


Figure 6. Geologic map along the balanced cross-section region (based on Mudge et al., 1982; Mudge and Earhart, 1983). Detail was added along the Sawtooth Range from several U.S. Geological Survey 1:24,000 quadrangle maps (Mudge, 1965, 1966a, 1966b, 1968), and our own mapping. S-S′ indicates position of cross section of Figure 7. Also shown is location of seismic line of Figure 8, additional seismic data used to construct the balanced cross section, and location of U-Pb sample KR2 of Figure 14. Location of map is shown in Figure 2. To view the full-sized PDF of Figure 6, please visit http://dx.doi.org/10.1130/GES00773.S2 or the full-text article at www.gsapubs.org.

the Marias River Shale and the Two Medicine Formation are evident in geological maps to the north and northwest of the Augusta syncline (Mudge and Earhart, 1983; Fig. 6). Closely imbricated Cretaceous rocks are restricted to shallow levels, as suggested by geological maps and seismic data.

The classic structures of the Sawtooth Range comprise a set of closely spaced, hinterland dipping thrust sheets involving Paleozoic to Lower Cretaceous sedimentary rocks (Mudge, 1982). The steep dip angles in these thrust sheets are a result of back rotation during the emplacement of successively younger and structurally lower thrust sheets to the east. The frontal structures detach at the base of the Devonian section, similar to the buried thrust sheets beneath the foothills. We infer that the basal detachment cuts downsection westward, because thrust sheets in the western part of the Sawtooth Range contain Cambrian strata (Figs. 6 and 7). Upper-level detachments are localized in the Ellis Group and Kootenai and Blackleaf Formations.

To the west of the Sawtooth Range, thrust slices containing Paleozoic rocks are buried by imbricated, highly deformed Mesozoic strata, and farther westward, by Belt Supergroup rocks in the hanging wall of the Lewis thrust system. Our interpretation of the subsurface geometry shows imbricated thrust faults repeating the Paleozoic section; this is required to avoid an excessive amount of shortening in the imbricated Cretaceous rocks, the regional elevation of which is ~3 km beneath the topographic surface. Moreover, the imbricate geometry fills space beneath the frontal part of the Lewis thrust system. The western extent of thrust sheets containing Paleozoic rocks is based on the interpretation of unpublished proprietary seismic reflection profiles shot across the northern part of the Continental Divide syncline and the South Fork of the Flathead River Basin (Figs. 2 and 7, points 21, 28). Although the seismic line that shows this relationship is located more than 30 km to the north, the continuity of surface structures along strike makes this projection permissible. The importance of the western extent of Paleozoic strata is that it provides an indirect but significant constraint for the possible location of the easternmost footwall ramp in Belt Supergroup strata. A conservative view is that this ramp may be located just westward of the footwall cutoff in the Paleozoic section, once thrust sheets of the Sawtooth Range and buried equivalents are restored to an undeformed position. The location of the frontal footwall cutoffs in Belt rocks, plus the unknown extent of eroded strata in thrust sheets of the Lewis thrust system, are the major sources of uncertainty in estimating shortening in this region.

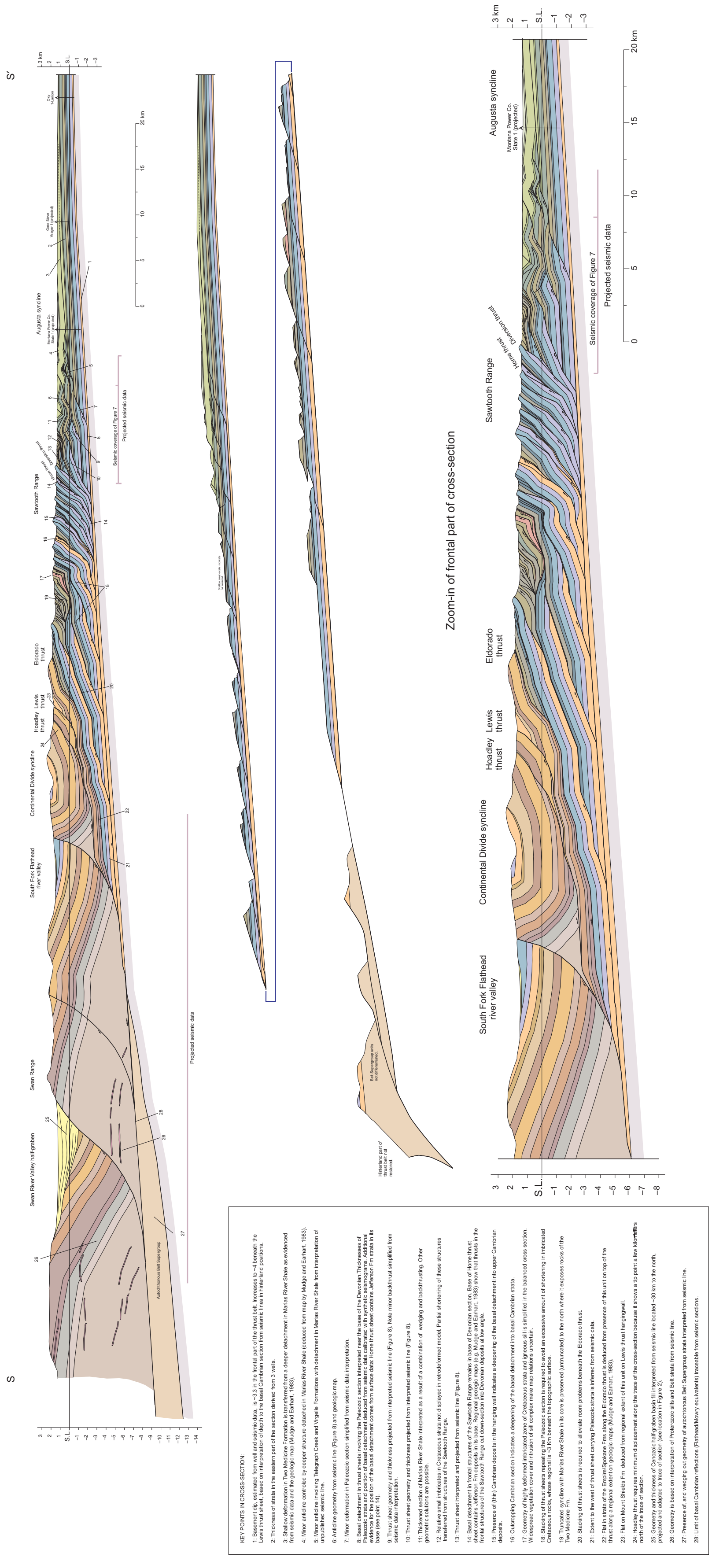


Figure 7. Balanced cross section. See Figure 6 for location and key for lithologic units. Western part of the cross section, west from the South Fork of the Flathead River Valley, is not shown in the restored section. To view the full-sized PDF of Figure 7, please visit http://dx.doi.org/10.1130/GES00773.S3 or the full-text article at www.gsapubs.org.

Stacks of thrust slices of Mesozoic strata occur on both sides of the Sawtooth Range. These thrusts propagated upward through the Paleozoic section from the basal detachment, were deflected eastward toward the foreland along detachments located low in the Mesozoic section, and were subsequently offset by underlying younger thrust faults. Because of its small scale and relative insignificance for calculating total shortening, the frontal imbricate east of the Diversion thrust is not shown in the restored section (Fig. 7). The detailed geometry in the similar imbricated system to the west of the Sawtooth Range is not strongly constrained, mainly because of the poor map resolution in this part of the thrust belt.

The frontal structure of the Lewis thrust system along the trace of the cross section is the Eldorado thrust. The traces of the Lewis and Eldorado thrusts diverge from a branch point located south of the Lewis thrust salient (Figs. 2 and 9). The Eldorado thrust follows a detachment in the Proterozoic Empire-Spokane Formations interval, as suggested by the long hanging-wall flat along these units shown in the Mudge and Earhart (1983) map. These two units are mapped as a single package at this position in the thrust belt because of their reduced thickness. In map view, the Lewis thrust also exhibits a relatively long hanging-wall flat in the Mount Shields Formation (Figs. 6 and 9). The Hoadley thrust has a relatively small amount of displacement in the cross section, consistent with the location of a tip point on the map a few kilometers north of the cross section (Figs. 6 and 9).

The Continental Divide syncline resulted partly from the juxtaposition of a hanging-wall ramp with a footwall flat beneath the western limb of the syncline (Fig. 7). Further tightening of the syncline took place by back rotation of the eastern limb of the fold during internal imbrication and additional passive tilting during emplacement of structurally lower thrust sheets in Paleozoic strata. The dip of bedding in the back limb is potentially accentuated by flexural slip, as suggested by the apparently high hanging-wall cutoff angle in lower Belt Supergroup units.

The western part of the cross section shows extensional faults produced by orogenic collapse and subsequent Basin and Range tectonics (Constenius, 1996). The location of the extensional fault just west of the Continental Divide syncline controls the position of the South Fork of the Flathead River. This fault probably accommodated the collapse of a structural culmination in a major fault-bend fold, which acquired additional structural elevation above superposed ramps along deeper thrust faults. The estimated normal slip along this fault is ~6 km.

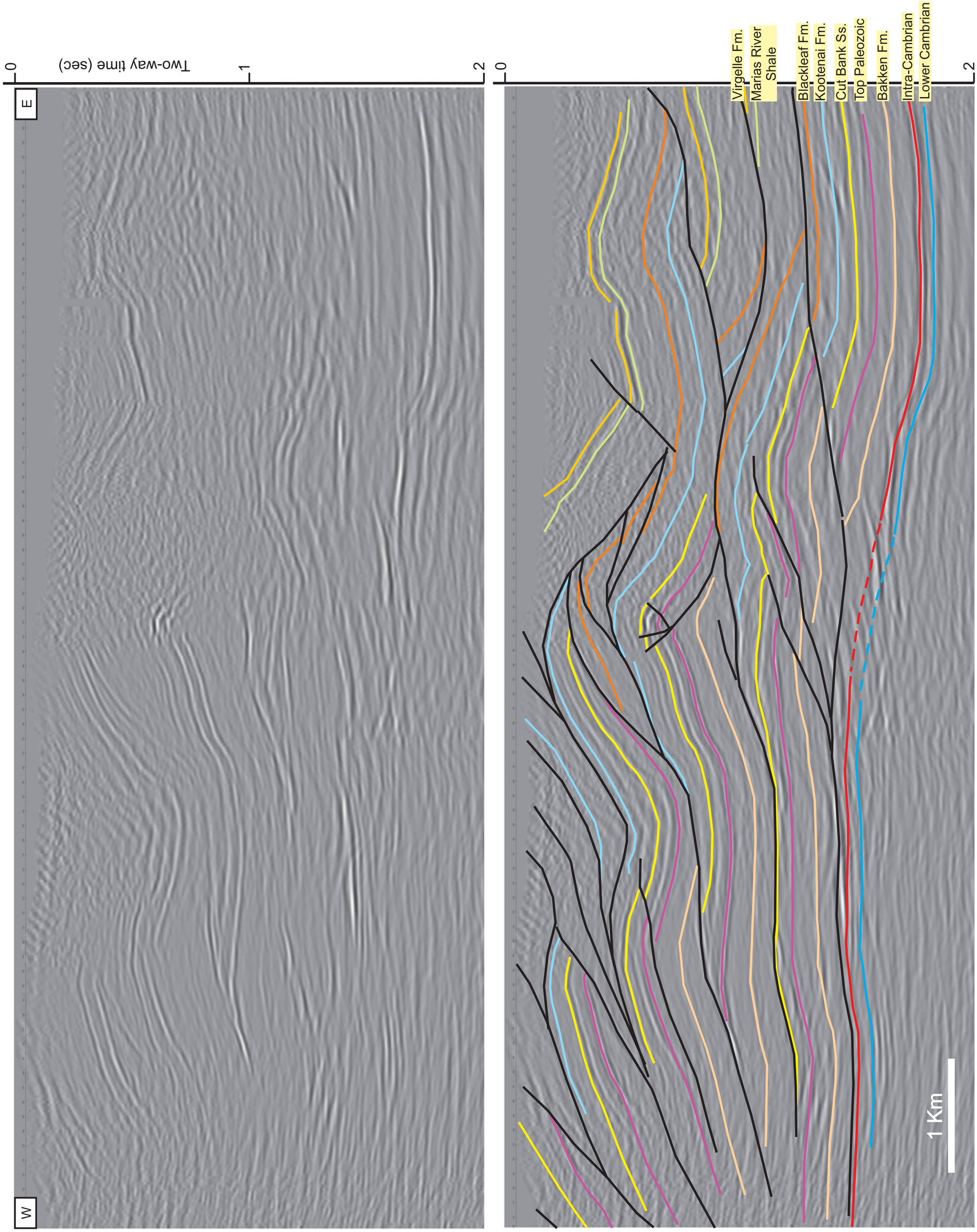


Figure 8. Uninterpreted and interpreted seismic section from the foothills, south of the Sun River. Vertical scale is two-way traveltime. Location is in Figure 6 (SL 1-a). A color version of this seismic section is available in Supplemental Figure 1 (see footnote 1).

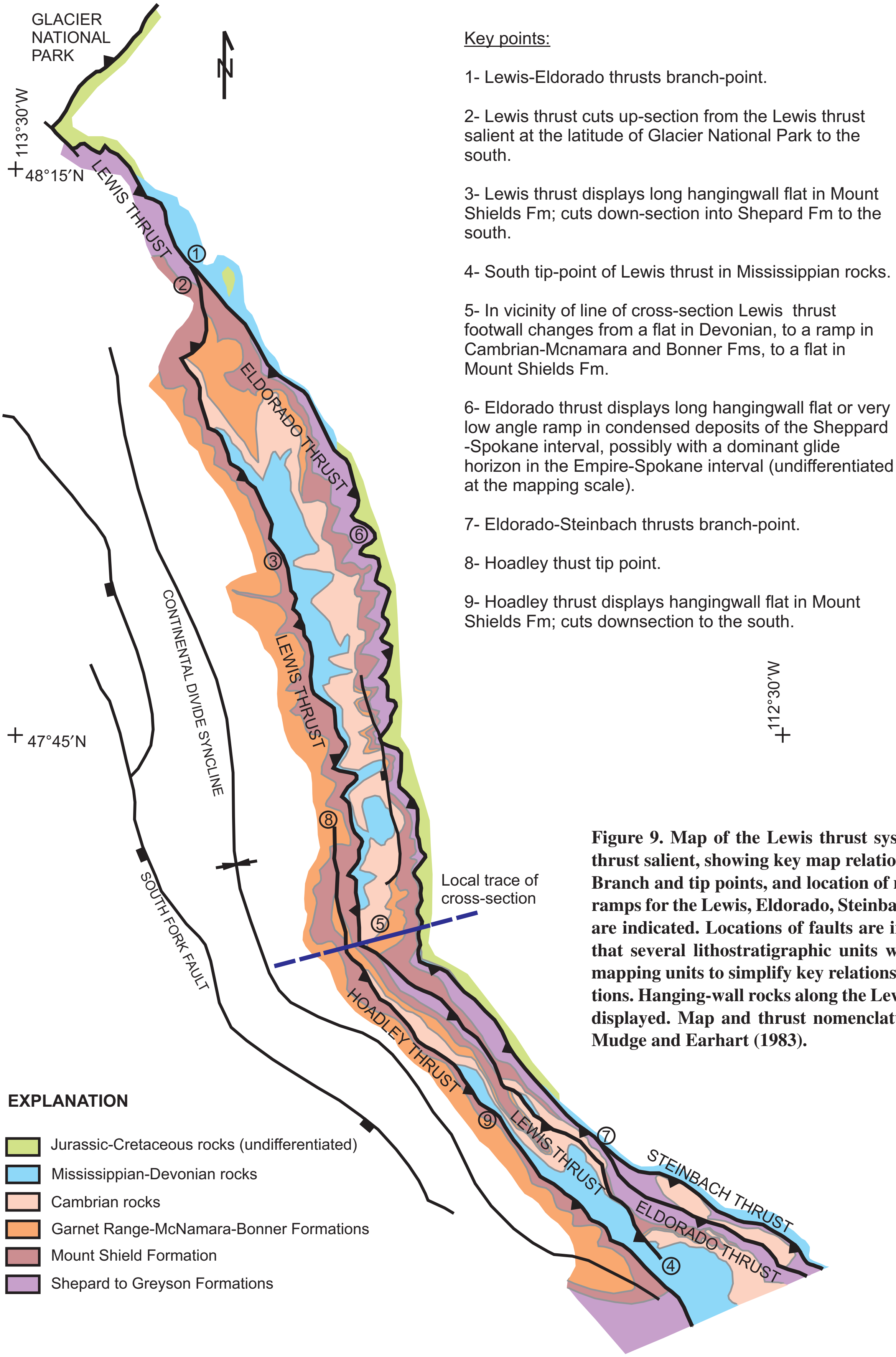


Figure 9. Map of the Lewis thrust system south of the Lewis thrust salient, showing key map relations of individual thrusts. Branch and tip points, and location of main glide horizons and ramps for the Lewis, Eldorado, Steinbach, and Hoadley thrusts are indicated. Locations of faults are in Figures 1 and 2. Note that several lithostratigraphic units were combined in single mapping units to simplify key relations and due to scale limitations. Hanging-wall rocks along the Lewis thrust salient are not displayed. Map and thrust nomenclature are simplified from Mudge and Earhart (1983).

The geometry of the major half-graben that controls the Swan River basin is partially based on projection of seismic data collected north of the plane of the cross section (Fig. 2). Thick mafic sills in Belt Supergroup strata generate strong seismic reflections that help to illustrate the structure. Depth estimates of extensional growth strata in the hanging wall of this structure indicate ~2.3 km of Cenozoic basin fill. Slip along the fault in the cross section is ~12 km. In addition, seismic data show stratified reflections beneath the basal Cambrian event that converge markedly toward the east. Our preferred interpretation is that these reflections correspond to autochthonous Belt Supergroup rocks tapering toward the east.

Total shortening along the cross section is ~135 km, or ~70% if measured from the axis of the Augusta syncline to the western flank of the Continental Divide syncline. The Lewis thrust system accommodates ~50 km of shortening along this cross section.

In southern Canada near the international border, van der Velden and Cook (1994) calculated ~115 km of total displacement for the Lewis thrust sheet by matching the Lewis thrust hanging wall and interpreted footwall cutoffs on a deep seismic reflection profile. Their estimate of shortening included the cumulative shortening of the Lewis thrust and all the other thrusts that affect the Paleozoic strata, and did not include additional shortening in western structures. Revision of the seismic profiles used in their study suggests that the location of the footwall ramp could be ~10 km farther west. Bally (1984) calculated a minimum of 165 km of shortening from a balanced cross section extending from the frontal part of the thrust belt in Canada to the Moyie thrust along the Idaho-Montana border. The cross section in Constenius (1996) indicates shortening of ~135 km, including the Lewis thrust system and all the structures to the east (Fig. 5). Based on these numbers, it appears that shortening remains relatively constant from the southern part of the Canadian thrust belt to the line of our cross section.

The main uncertainties on our shortening estimate are as follows. The geometry in the frontal part of the section is well constrained by seismic data; minor variations in the interpretation are possible but could not significantly affect the shortening amount. The structure in the Sawtooth Range does not permit many geometric solutions, and the solutions chosen in the cross section represent minimum shortening. The detailed geometry of the partially buried stack of thrust slices of Paleozoic strata between the Sawtooth Range and the Lewis thrust system is conjectural owing to the lack of subsurface data. These repetitions add shortening, but are required to fill space as discussed herein. The extent of the westernmost thrust sheet of Paleozoic rocks beneath the Lewis thrust system adds a few kilometers of shortening and is inferred from relatively ambiguous seismic data. The largest sources of uncertainty in the shortening estimate are the westward extent of these buried thrust sheets, the location of the footwall ramp in Belt Supergroup strata, and the length of eroded strata in the leading edge of the Lewis thrust system. The position of the basal detachment is well constrained by subsurface data and cannot be brought to shallower levels to decrease shortening.

The location of the footwall cutoffs in Belt Supergroup strata derived from the cross-section restoration coincide with the inferred positions of the footwall megaramp of Price and Sears (2000) and Sears (2001, 2007) at this latitude in the thrust belt; shortening may decrease southward from this position, as suggested by these authors.

The transition from the deeply exhumed Sawtooth Range, in the south, to the less exhumed Lewis thrust salient with its thick Belt Supergroup section to the north (Figs. 2, 5, and 10) results from two factors: (1) regional basement elevation increases from north to south, as indicated by seismic profiles, well data, and balanced cross sections; and (2) the Flathead and Waterton duplexes beneath the Lewis thrust salient apparently merge southward into, or are replaced by, a single large exhumed duplex in the Sawtooth Range. Consequently, structural elevations of Paleozoic strata in thrust sheets in the Sawtooth Range reach elevations more than 2 km above sea level. In contrast, the top of the Paleozoic section between the Flathead and Waterton duplexes, beneath the Akamina syncline, is ~3.5–3.6 km below sea level (Gordy et al., 1977; Bally, 1984). A possible third factor is that the Lewis thrust sheet salient is bounded on its northwest and southeast sides by lateral ramps, across which the Lewis thrust climbs upsection along its hanging wall from a detachment located deep in the Prichard Formation (or lateral equivalents) into much higher stratigraphic levels. The Lewis thrust sheet is thicker and stronger within the salient than in the flanking regions. North of Marias Pass the Lewis thrust sheet includes a thick succession of the Prichard and Appekunny Formations that is absent along the hanging wall of the Lewis thrust system south of Marias Pass.

KINEMATIC HISTORY

A preliminary kinematic history of the Cordilleran thrust belt at the latitude of northwestern Montana can be deduced from crosscutting and overlapping relationships between rock units of known ages, provenance of dated synorogenic sedimentary units in the foreland basin system, and previously published apatite fission-track ages (Fig. 11).

Early Hinterland Deformation (Jurassic)

By Jurassic time, shortening in hinterland regions of the eastern part of the Intermontane and Omineca belts was driven by the collision of the Intermontane terrane with North America (Evenchick et al., 2007, and references therein). In Canada, accretion of the Intermontane terrane between 187 and 173 Ma involved tectonic burial of the outboard part of the Cordilleran miogeocline to depths of 20–25 km, and was accompanied by regional metamorphism and the emplacement of synkinematic granitic plutons. Subsequent southwest-verging deformation of the outboard part of the miogeocline involved ~10 km of rapid exhumation between 173 and 168 Ma (Archibald et al., 1983; Colpron et al., 1996, 1998; Evenchick et al., 2007).

Evidence of hinterland deformation at the latitude of northwestern Montana is deduced from modal petrographic data and U-Pb detrital zircon ages from Jurassic sandstones (Fuentes et al., 2009, 2011). Ellis Group and Morrison Formation sandstones contain lithic fragments that suggest derivation from deformed miogeoclinal strata with some influence from volcanic sources (Fig. 12; Table 1). The key information that proves a western, orogenic provenance is found in detrital zircon ages. Samples of the Swift and Morrison Formations contain abundant zircons with late Paleozoic and Triassic ages derived from accreted elements in the eastern part of the Intermontane belt, mid-Paleozoic ages with possible origin in Kootenay arc rocks, and a variety of different miogeoclinal source units (Fig. 13; Supplemental Table[2]). This interpretation is consistent with the Late Jurassic onset of foreland basin subsidence recorded by ~1000 m of Kootenay Group deposits that crop out near Fernie, British Columbia (Ross et al., 2005).

Western Thrust Systems (Cretaceous)

Basal coarse-grained sandstone and conglomerate of the Lower Cretaceous Kootenai Formation and equivalent strata in southwestern Montana, Wyoming, and southern Canada contain a high proportion of chert, which has

[2]Supplemental Table. Excel file of U-Pb (zircon) geochronologic analyses by laser-ablation multi-collector–inductively coupled plasma–mass spectrometry. If you are viewing the PDF of this paper or reading it offline, please visit http://dx.doi.org/10.1130/GES00773.S4 or the full-text article on www.gsapubs.org to view the Supplemental Table.

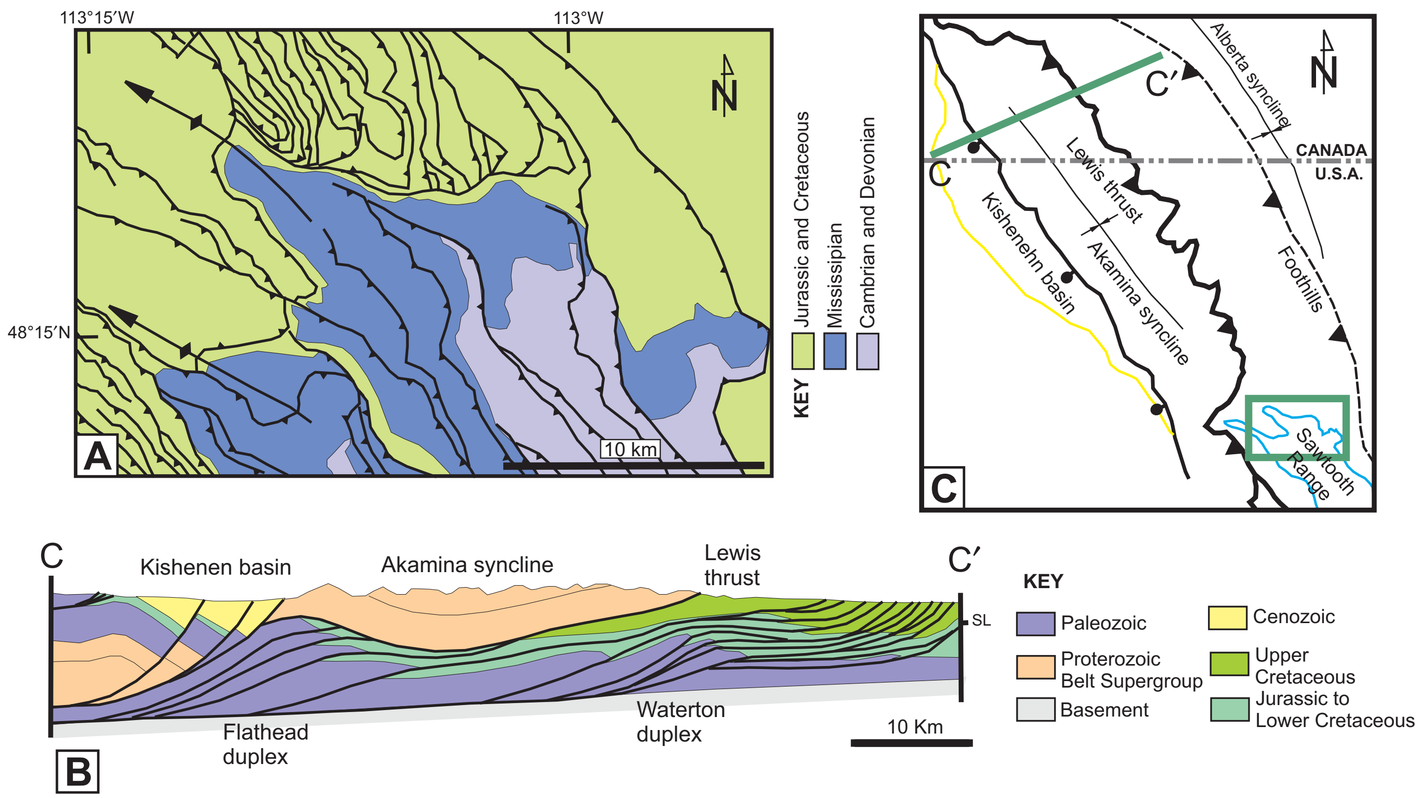


Figure 10. (A) Simplified geologic map of the northern end of the Sawtooth Range showing the divergence of the single imbricate fan into two plunging individual systems to the north. Map is based on Mudge and Earhart (1983). (B) Simplified cross section showing the Waterton and Flathead duplexes and their relation with the Akamina syncline in proximity of the international border. Based on Gordy et al. (1977). (C) Relative location map showing trace of cross section and detailed map (A).

long been interpreted to reflect derivation from mid- to upper Paleozoic rocks that were uplifted and eroded in the growing thrust belt to the west (Rapson, 1965; Suttner et al., 1981; DeCelles, 1986; Schwartz and DeCelles, 1988). The high ratio of sedimentary lithic to total lithic fragments (Fig. 12), together with the high proportion of detrital zircons with ages typical of miogeoclinal strata, indicates a provenance dominated by thrust-imbricated Paleozoic rocks. Uncertainty exists regarding the location of this early fold-thrust belt; the locus of deformation may have resided in the region occupied today by the Omineca belt, or along the western flank of the Purcell anticlinorium.

The oldest known structures in the unmetamorphosed western part of the thrust belt are the Hall Lake and St. Mary thrusts, and the Lussier River fault, in the vicinity of the United States–Canada border. Palinspastic map reconstructions by Price and Sears (2000) indicate that part of the Hall Lake and St. Mary thrusts were located mainly south of the international border prior to the movement of thrust systems to the east. The timing of initial slip along these structures is unknown, but mid-Cenomanian crosscutting plutons indicate pre–mid-Cretaceous movement (Wanless et al., 1968; Foo, 1979; Archibald et al., 1984; Höy and van der Heyden, 1988, Price and Sears, 2000). Recent work in southern British Columbia provides additional information about the timing of early contractional deformation in this region: $^{40}Ar/^{39}Ar$ dating of stocks intruding and thermally overprinting the Lussier River fault and deformed Cambrian–Ordovician shales and carbonate rocks yielded ages of ca. 108 Ma (Larson et al., 2006). All these geochronometric results in southern Canada fix the minimum age of displacement on the St. Mary–Lussier fault system between mid-Albian and late Cenomanian time.

In northern Idaho the Early and Middle Eocene extensional exhumation of the Priest River metamorphic core complex from depths of 35–40 km removed the thrust belt, locally exposing its basal detachment and the underlying Archean basement (Doughty et al., 1998; Doughty and Price, 1999). Although the preextensional structural configuration of this region is uncertain, thrusts involving Belt Supergroup and younger rocks west of the Priest River complex have been identified (Rhodes and Hyndman, 1988; Stoffel et al., 1991). Early Cretaceous deformation involving Belt Supergroup rocks in hinterland regions seems to have been a regional process, as indicated by an abrupt increase in the proportion of low-grade metamorphic lithic fragments and detrital zircons in the Blackleaf Formation with U-Pb ages typical of Belt Supergroup strata (Figs. 12 and 13; Supplemental Table [see footnote 2]) (Fuentes et al., 2011).

The Moyie thrust system extends from British Columbia into northwestern Montana (Figs. 1 and 2). Shortening estimates range from ~8 km in southeastern British Columbia (McMechan and Price, 1982) to ~4 km in the central Cabinet Mountains of Montana (Fillipone and Yin, 1994). Slip along this thrust has been bracketed between ca. 71 and 69 Ma (Fillipone and Yin, 1994), based on $^{40}Ar/^{39}Ar$ ages of mylonitic rocks from the west side of the Dry Creek stock (Fig. 2) that were interpreted to be deformed by the Moyie thrust. This interpretation, however, can only limit the timing of the youngest slip along this thrust. Approximately 50 km to the south, a ca. 114 Ma $^{40}Ar/^{39}Ar$ hornblende age from the Vermillion River stock, which was interpreted to crosscut the Moyie

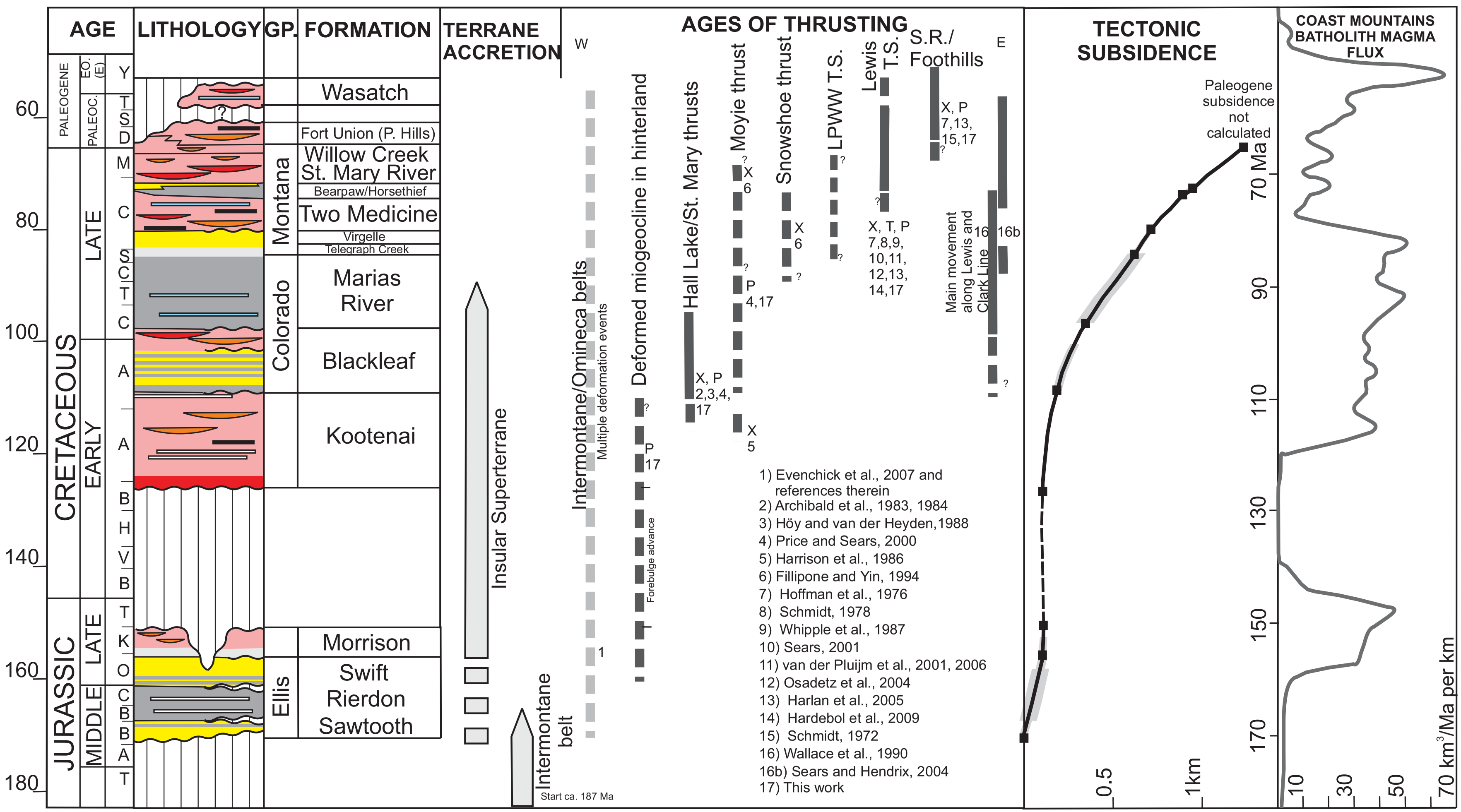


Figure 11. Stratigraphic chart, major deformation events in the fold-thrust belt, and tectonic subsidence of the foreland basin system at the latitude of northern Montana. Sources of data for timing of thrust movement numbered and given in diagram. Timing of terranes accretion discussed in the text. SLPWW T.S.—Snowshoe-Libby-Pinkham-Whitefish-Wigwam thrust system; Lewis T.S.—Lewis-Eldorado-Steinbach-Hoadley thrust system; S.R.—Sawtooth Range; P. Hills—Porcupine Hills. Time constraints of thrust systems: X—crosscutting relationships; P—provenance data; T—thermochronology (apatite fission tracks). Subsidence curve from Fuentes et al. (2009, 2011). Timing of deformation in hinterland regions of the Intermontane and Omineca belts at the latitude of Montana is unknown, but it is constrained in southern Canada. Coast Mountains batholith magmatic flux is from Gehrels et al. (2009).

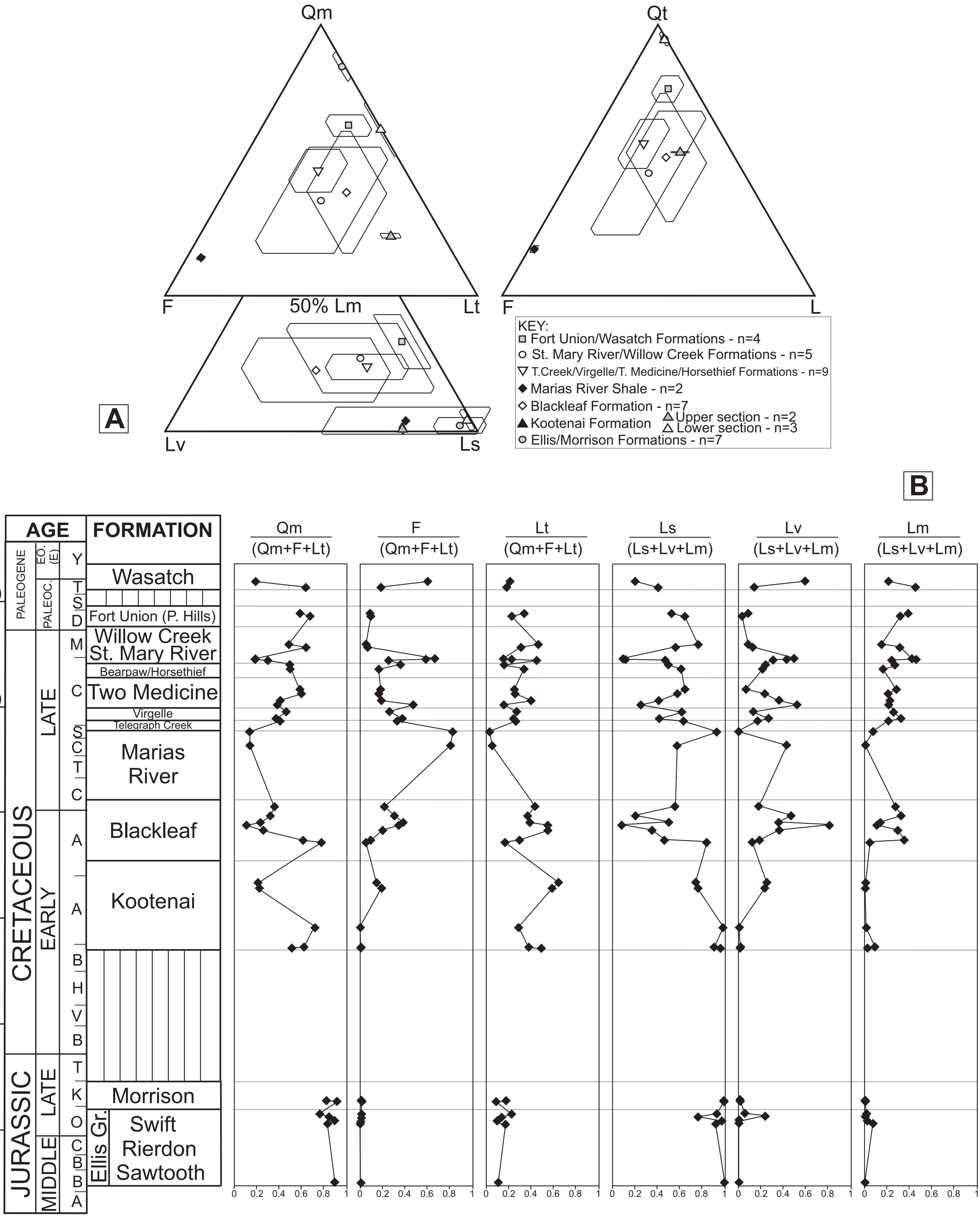


Figure 12. (A) Ternary diagrams illustrating mean modal framework-grain compositions of foreland basin sandstones with standard deviations. Qm—monocrystalline quartz; Qt—total quartz; Lt—total lithics; F—feldspar; Lv—volcanic lithics; Ls—sedimentary lithics. (B) Plot showing the stratigraphic distribution of ratios of main grain types. For explanation of petrographic parameters, see Table 1. Both figures after Fuentes et al. (2011).

TABLE 1. PETROGRAPHIC PARAMETERS

Quartzose grains	
Qm	Monocrystalline quartz
Qp	Polycrystalline quartz
Qpt	Foliated polycrystalline quartz
Qss	Quartz in sandstone or quartzite lithic grain
S	Siltstone
Qt	Total quartzose grains (Qm + Qp + Qpt + Qss + S + C + Cb)
Feldspar grains	
K	Potassium feldspar
P	Plagioclase feldspar
F	Total feldspar grains (K + P)
Lithic grains	
Metamorphic	
Lph	Phyllite
Lsm	Schist
Lss	Serpentine schist
M	Marble
Lm	Metamorphic lithic grains (Lph + Lsm + Lss + M + Qpt)
Volcanic	
Lvl	Lathwork volcanic grains
Lvx	Microlitic volcanic grains
Lvf	Felsic volcanic grains
Lvv	Vitric volcanic grains
Lvm	Mafic volcanic grains
Lv	Total volcanic lithic grains (Lvl + Lvx + Lvf + Lvv + Lvm)
Sedimentary	
D	Dolostone
Lc	Limestone
Lsh	Shale or mudstone
C	Chert
Cb	Black chert
Ls t	Total sedimentary lithic grains (D + Lc + Lsh + C + Cb + Qss + S)
Lt	Total lithic grains (Ls + Lv + Lm)
L	Unstable lithic fragments [Lt (C + Cb + Qss + S + Qpt)]

Note: Accessory minerals: epidote, zoisite, chlorite, garnet, amphibole, muscovite, biotite, tourmaline, pyroxene, kyanite, cordierite, sillimanite, zircon, kaolinite, glauconite, and apatite.

thrust (Harrison et al., 1986), suggests that this fault was active earlier than Late Cretaceous time. Possible explanations for this discrepancy include: (1) the lack of clear crosscutting relations led to erroneous interpretations; and (2) the Vermillion River stock cuts a splay of the Moyie system different from the one that cuts the Dry Creek stock. If this were the case, it would be possible that movement on this thrust system occurred during pre–late Aptian time, with selective reactivation of splays during the late Campanian–Maastrichtian. This view aligns with the suggestion of Early Cretaceous or older slip along the Moyie thrust (Price and Sears, 2000) based on structural similarities among the Moyie, St. Mary, and Hall Lake faults, and the recent age data from Larson et al. (2006) discussed herein.

The Snowshoe thrust, mapped as a forethrust by Harrison et al. (1992), was reinterpreted as an east-dipping backthrust system by Fillipone and Yin (1994). The timing of slip on this thrust system is unknown. Based on local juxtaposition of west-dipping cleavage in the Moyie thrust sheet over east-dipping cleavage related to structures of the Snowshoe thrust system, and truncation of west-vergent folds by the Moyie thrust, Fillipone and Yin (1994) suggested that the Moyie thrust postdates the Snowshoe thrust system.

Uncertainty also exists for the timing of slip along the Libby and Pinkham thrust systems, as well as other thrusts to the east, mapped as the Whitefish and Wigwam thrusts by Harrison et al. (1992). Provenance data do not provide constraints for the timing of displacement on these faults because of the similar hanging-wall stratigraphy of the thrust sheets involved. All of these thrusts sheets detach in and expose Belt Supergroup strata, and prior to exhumation and erosion contained miogeocline rocks and possibly early foreland basin deposits.

Lewis Thrust System (Late Cretaceous–Early Paleogene)

Kinematic history information is better for the eastern part of the thrust belt (Fig. 11). Apatite and zircon fission track dating in the Lewis thrust salient in southern Canada suggest rapid hanging-wall cooling ca. 75 Ma (Osadetz et al., 2004; Feinstein et al., 2007). The $^{40}Ar/^{39}Ar$ dating of clay-bearing gouge (illite) led van der Pluijm et al. (2006) to suggest pulses of contractional deformation for the Canadian segment of the Lewis thrust ca. 72 and ca. 52 Ma. Forward thermokinematic modeling by Hardebol et al. (2009) is in agreement with onset of deformation ca. 75 Ma. None of these data exclude the possibility of earlier displacement. The apatite and zircon cooling ages could be a result of passive uplift during duplex development beneath the Lewis thrust. Dates from fault gouge at isolated localities may represent discrete episodes of movement, especially considering that van der Pluijm et al. (2006) showed two different age clusters separated by ~20 m.y. The illite ages could also be related to diagenetic processes in fault zones, rather than kinematic events. Price (2007) and Pevear et al. (2007) presented an interesting discussion about the validity of the van der Pluijm et al. (2006) ages.

Sears (2001) proposed that the Lewis thrust in Montana did not move until 74 Ma, based on two lines of argument: (1) correlative ca. 76 Ma sills intruded undeformed Cretaceous rocks in both footwall and hanging-wall positions prior to thrusting; and (2) 74 Ma ash-fall deposits covered footwall and hanging-wall rocks of the Lewis thrust sheet. Andesite sills in the hanging wall yielded a biotite $^{40}Ar/^{39}Ar$ age of 75.9 ± 1.2 Ma (Sears et al., 1997, 2000). However, these sills are restricted to the Garrison depression, located southwest of the Lewis thrust sheet, in the domain of the Lewis and Clark line and Helena salient (Sears et al., 2000). The assumed correlative sills in the Lewis thrust system footwall intruded concordant to bedding in the Kootenai and Blackleaf Formations; these sills are now exposed in strongly deformed thrust sheets to the west of the Sawtooth Range (Fig. 6). A sample of one sill yielded a biotite $^{40}Ar/^{39}Ar$ age of 58.8 ± 1.5 Ma, with a disturbed gas release spectrum, and a sericite $^{40}Ar/^{39}Ar$ age of ca. 60 Ma (Sears et al., 2000; Sears, 2001); these ages were interpreted as cooling ages (Sears, 2001). New U-Pb geochronology of zircons from a sill sample collected west of Gibson Reservoir (Fig. 6) was conducted by laser ablation–multicollector–inductively coupled plasma mass spectrometry (LA-MC-ICPMS) at the Arizona LaserChron Center. The analytical data are reported in Table 2. A weighted mean age of 82.8 ± 0.8 Ma for this sill sample (Fig. 14) shows that not all of the sills in the Cretaceous rocks cut by the Lewis thrust can be correlated confidently. The 74 Ma ash could have been deposited on top of moving thrust sheets. Although the proposed onset of slip along the Lewis thrust by Sears (2001) is coincident with timing deduced from work in Canada, we suggest that earlier episodes of slip along this structure cannot be ruled out. The new U-Pb age provides a ca. 83 Ma limit for

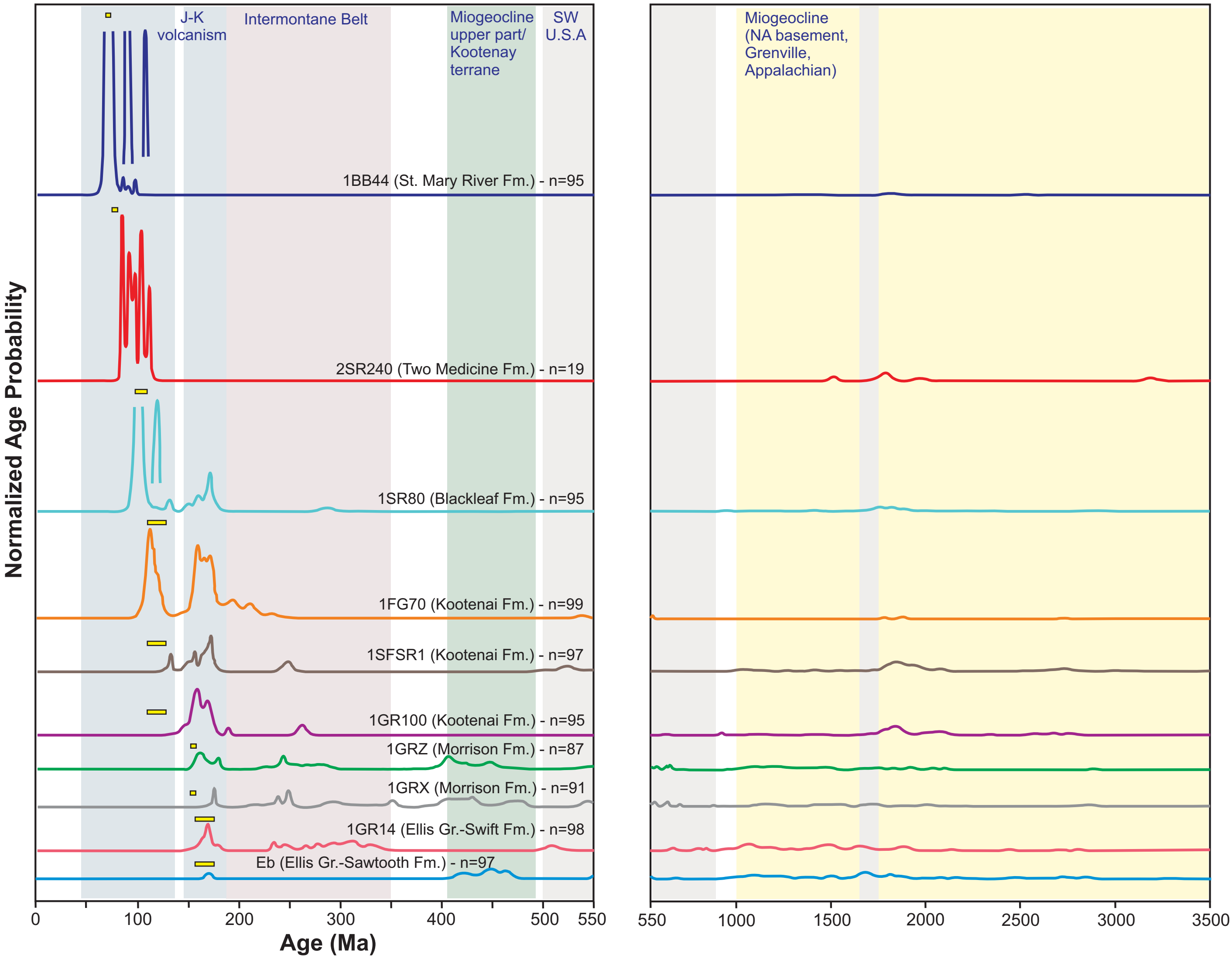


Figure 13. Age probability plots of U-Pb ages of detrital zircons of selected samples from foreland basin deposits. Shaded areas indicate main origins of zircons (from Coney and Evenchick, 1994; Roback and Walker, 1995; Gehrels and Ross, 1998; Ross and Villeneuve, 2003; Ross et al., 2005; Link et al., 2007). Yellow bars indicate depositional age of samples. Note change in horizontal scale in plots at 550 Ma. Modified from Fuentes et al. (2011). Data are in Supplemental Table (see footnote 2).

the maximum age of deformation of Cretaceous rocks east of the Lewis thrust system.

Key constraints on the timing of displacements on the Lewis thrust system and the frontal structures are provided by sills and dikes associated with the Adel Mountain Volcanics, in the southern part of this segment of the thrust belt. In the Rogers Pass area (Fig. 15), a monzonite intrusion postdates motion along the Steinbach thrust (Whipple et al., 1987; Mudge et al., 1982; Harlan et al., 2005). Although the intrusion has a complex geometry, it clearly cuts across the trace of the fault (Fig. 15). A K-Ar age (biotite) of 59.6 ± 1.6 Ma (Schmidt, 1978; Whipple et al., 1987) has been used to establish the minimum possible age of movement along this thrust system (Sears, 2001). New U-Pb ages were obtained from 20 zircon crystals from a sample of this intrusive at the University of Arizona LaserChron Center. A weighted mean age of 52.6 ± 0.4 Ma limits the youngest possible age of slip along this fault to the Early Eocene (Fig. 15). To the east, a number of monzonite porphyry sills intruded into Cretaceous rocks are folded and cut by minor thrusts, indicating that sill emplacement predated or was synchronous with shortening (Schmidt, 1972). The K/Ar and $^{40}Ar/^{39}Ar$ ages of these sills (Harlan et al., 2005) suggest thrusting younger than 55.5 Ma.

The new U-Pb ages bracket the youngest movement on the Steinbach thrust; however, the termination of slip on the Lewis thrust system has not been precisely dated because the thrust sheets have been deeply eroded along their leading edges, removing any Late Cretaceous–Early Eocene strata that may have overlain or been truncated by these faults. The youngest unit truncated by the Lewis thrust system is the Campanian Two Medicine Formation. An increase in low-grade metamorphic grains in sandstones of the lower part of the late Campanian–Maastrichtian St. Mary River Formation (Fig. 12) possibly reflects exhumation of Belt Supergroup rocks above this thrust system. A similar increase in low-grade metamorphic detritus at equivalent stratigraphic levels was noted by Mack and Jerzykiewicz (1989) in southern Canada.

TABLE 2. U-Pb GEOCHRONOLOGIC ANALYSES

				Isotope ratios					Apparent ages (Ma)					
Analysis	$^{206}Pb/^{204}Pb$	$^{206}Pb^*/^{207}Pb^*$	± (%)	$^{207}Pb^*/^{235}U^*$	± (%)	$^{206}Pb^*/^{238}U$	± (%)	error correction	$^{206}Pb^*/^{238}U^*$	± (Ma)	$^{207}Pb^*/^{235}U$	± (Ma)	$^{206}Pb^*/^{207}Pb^*$	± (Ma)
2RP-1	8364	16.6804	13.6	0.0689	14.3	0.0083	4.3	0.30	53.5	2.3	67.6	9.4	601.8	296.2
2RP-2	14980	17.4316	10.3	0.0624	11.2	0.0079	4.4	0.39	50.6	2.2	61.4	6.7	505.7	226.6
2RP-3	7588	23.2611	22.1	0.0487	22.2	0.0082	1.8	0.08	52.7	1.0	48.3	10.5	–168.2	556.9
2RP-5	16788	25.8982	85.2	0.0434	85.2	0.0082	1.3	0.02	52.4	0.7	43.2	36.0	–442.9	2851.5
2RP-6	113360	24.7918	15.3	0.0475	15.4	0.0085	1.3	0.08	54.8	0.7	47.1	7.1	–329.4	395.8
2RP-7	16560	24.5352	28.5	0.0455	28.6	0.0081	1.8	0.06	52.0	0.9	45.2	12.6	–302.7	742.5
2RP-8	20760	18.1797	24.1	0.0621	24.2	0.0082	2.2	0.09	52.5	1.2	61.1	14.4	412.5	545.8
2RP-9	6404	18.2371	45.1	0.0625	45.1	0.0083	2.2	0.05	53.1	1.1	61.6	27.0	405.4	1060.5
2RP-10	14448	24.7605	20.8	0.0474	20.8	0.0085	1.6	0.08	54.6	0.9	47.0	9.6	–326.1	537.9
2RP-11	15944	20.1302	28.4	0.0617	28.4	0.0090	1.7	0.06	57.8	1.0	60.8	16.8	179.9	673.6
2RP-12	5336	12.9089	55.4	0.0907	55.5	0.0085	2.9	0.05	54.5	1.6	88.1	46.9	1133.2	1202.2
2RP-13	13512	22.5989	7.7	0.0509	8.3	0.0083	3.0	0.36	53.6	1.6	50.4	4.1	–96.7	190.1
2RP-14	9644	18.4367	25.7	0.0636	26.0	0.0085	4.0	0.15	54.6	2.2	62.6	15.8	381.0	585.7
2RP-15	14536	20.4294	28.1	0.0536	28.1	0.0079	1.5	0.05	51.0	0.8	53.0	14.5	145.4	670.1
2RP-16	85040	22.0611	16.6	0.0505	16.7	0.0081	1.4	0.08	51.9	0.7	50.0	8.2	–37.9	406.4
2RP-17	27588	22.6283	11.1	0.0499	11.1	0.0082	1.0	0.09	52.6	0.5	49.5	5.4	–99.9	272.4
2RP-18	33244	19.8337	9.8	0.0547	10.2	0.0079	2.8	0.27	50.6	1.4	54.1	5.4	214.4	227.2
2RP-19	10376	19.7461	14.8	0.0565	14.9	0.0081	2.3	0.15	52.0	1.2	55.8	8.1	224.6	342.8
2RP-20	25084	23.4852	11.5	0.0458	11.7	0.0078	2.0	0.17	50.0	1.0	45.4	5.2	–192.1	287.9
2RP-21	2252	13.7260	23.8	0.0851	23.9	0.0085	2.7	0.11	54.4	1.5	83.0	19.1	1009.9	489.0
KR2-1C	5986	18.6375	23.0	0.0891	23.8	0.0120	5.8	0.24	77.2	4.4	86.7	19.7	356.6	526.5
KR2-2C	9985	18.9472	15.2	0.0902	15.7	0.0124	3.9	0.25	79.4	3.1	87.7	13.2	319.3	347.5
KR2-3C	14415	21.0558	1.2	0.0817	2.8	0.0125	2.5	0.89	79.9	2.0	79.7	2.1	74.1	29.7
KR2-4C	18273	21.1273	1.7	0.0842	3.1	0.0129	2.6	0.84	82.6	2.1	82.1	2.4	66.0	39.6
KR2-5R	17771	20.6464	3.3	0.0860	3.8	0.0129	1.9	0.49	82.5	1.5	83.7	3.1	120.6	78.3
KR2-6R	21552	20.9000	2.2	0.0831	2.8	0.0126	1.8	0.64	80.7	1.4	81.0	2.2	91.7	51.0
KR2-8	17434	21.2500	7.5	0.0861	8.1	0.0133	3.0	0.37	85.0	2.5	83.9	6.5	52.2	179.5
KR2-9	19930	21.8703	3.2	0.0798	4.3	0.0127	2.9	0.68	81.1	2.4	78.0	3.3	–16.9	77.3
KR2-10	15117	16.6684	27.1	0.1081	27.1	0.0131	2.2	0.08	83.7	1.8	104.2	26.9	603.3	595.6
KR2-11	30097	21.0068	3.6	0.0870	4.1	0.0133	1.9	0.47	84.9	1.6	84.7	3.3	79.6	85.9
KR2-12	20500	20.8042	3.8	0.0862	4.8	0.0130	2.9	0.60	83.3	2.4	83.9	3.9	102.6	91.0
KR2-13	18877	21.0911	5.0	0.0862	5.2	0.0132	1.3	0.25	84.5	1.1	84.0	4.2	70.1	118.9
KR2-14	13490	21.6132	5.0	0.0758	10.0	0.0119	8.6	0.87	76.1	6.5	74.2	7.1	11.6	120.3
KR2-15	38379	20.8751	3.3	0.0857	3.9	0.0130	1.9	0.50	83.1	1.6	83.5	3.1	94.5	79.2
KR2-16	23194	20.4416	5.5	0.0873	6.0	0.0129	2.3	0.39	82.9	1.9	85.0	4.9	144.0	129.5
KR2-18	21508	20.1581	4.2	0.0875	4.4	0.0128	1.5	0.34	81.9	1.2	85.2	3.6	176.6	97.0
KR2-19	16619	20.2485	2.4	0.0891	3.6	0.0131	2.6	0.74	83.8	2.2	86.6	3.0	166.2	56.6
KR2-20	17869	22.4863	4.7	0.0785	5.6	0.0128	3.1	0.56	82.0	2.6	76.8	4.2	–84.5	114.0
KR2-21	19908	20.8994	5.5	0.0811	6.8	0.0123	4.0	0.59	78.8	3.2	79.2	5.2	91.8	129.7
KR2-22	2613	17.8995	13.8	0.0993	14.1	0.0129	2.6	0.18	82.5	2.1	96.1	12.9	447.1	308.7
KR2-23	13805	18.8866	20.5	0.0959	20.6	0.0131	2.1	0.10	84.1	1.8	92.9	18.3	326.5	469.3
KR2-24	27896	20.9723	2.7	0.0871	3.6	0.0133	2.3	0.66	84.9	2.0	84.8	2.9	83.5	63.6

Note: All uncertainties are reported at the 1σ level, and include only measurement errors. Systematic errors would increase age uncertainties by 1%–2%. U concentrations and U/Th are calibrated relative to our Sri Lanka zircon standard, and are accurate to ~20%. Common Pb correction is from ^{204}Pb, with composition interpreted from Stacey and Kramers (1975) and uncertainties of 1.0 for $^{206}Pb/^{204}Pb$, 0.3 for $^{207}Pb/^{204}Pb$, and 2.0 for $^{208}Pb/^{204}Pb$. U/Pb and $^{206}Pb/^{207}Pb$ fractionation is calibrated relative to fragments of a large Sri Lanka zircon of 564 ± 4 Ma (2σ). U decay constants and composition as follows: ^{238}U = 9.8485 x 10 – 10, ^{235}U = 1.55125 x 10 – 10, $^{238}U/^{235}U$ = 137.88.

Sawtooth Range and Foothills (Late Cretaceous?–Early Paleogene)

Thrust displacement along the structures of the Sawtooth Range postdates, at least in part, displacement on the Lewis thrust. Map relations and cross sections along the Lewis thrust salient show that the Akamina syncline resulted from passive folding during the emplacement of the Waterton and Flathead duplexes (Bally et al., 1966; Gordy et al., 1977; Fermor and Moffat, 1992; Mudge and Earhart, 1983). These two duplexes are the northern equivalent of the Sawtooth Range (Fig. 10).

Samples from Cretaceous bentonite beds in thrust sheets of the Sawtooth Range and foothills yielded K/Ar ages (illite-smectite) of between 72 and 56 Ma that were interpreted as the results of low-grade burial metamorphism from stacking of thrust sheets (Hoffman et al., 1976). As Lageson (1987) pointed out, this age range does not necessarily indicate the beginning and end of thrusting; instead it represents the time span during which maximum temperatures were reached, which may correspond to thrust emplacement. Moreover, the K-Ar ages for bentonites provide crude constraints, but detailed interpretations are problematic as relative contributions from tectonic and sedimentary burial, potential effects of fluids, and potential problems from incomplete resetting and mixing of small detrital components are difficult to evaluate. Hoffman et al. (1976) and Hoffman and Hower (1979) assumed that the maximum thickness of strata above the stratigraphic interval sampled (Blackleaf Formation to Two Medicine Formation) was 350–1350 m; this requires a considerable tectonic thickening to reach the 100–175 °C indicated by the low-grade metamorphic mineralization. Recent thermo-kinematic work in southern Canada indicates that a thickness of 3–5 km of synorogenic strata was removed from proximal positions (Hardebol et al., 2009), calling into question the interpretation of burial metamorphism driven solely by thrust sheet burial. In any case, the timing proposed by Hoffman et al. (1976) agrees with the general timing of structure formation in the frontal part of the thrust belt.

The youngest sedimentary unit truncated by thrusts in the foothills region is the Willow Creek Formation (Mudge and Earhart, 1983), which is broadly dated by vertebrate and invertebrate fossils as Maastrichtian–Early Paleocene (Russell, 1950, 1968; Tozer, 1956; Catuneanu and Sweet, 1999). Along the triangle zone of the Alberta syncline, rocks of the early Paleocene

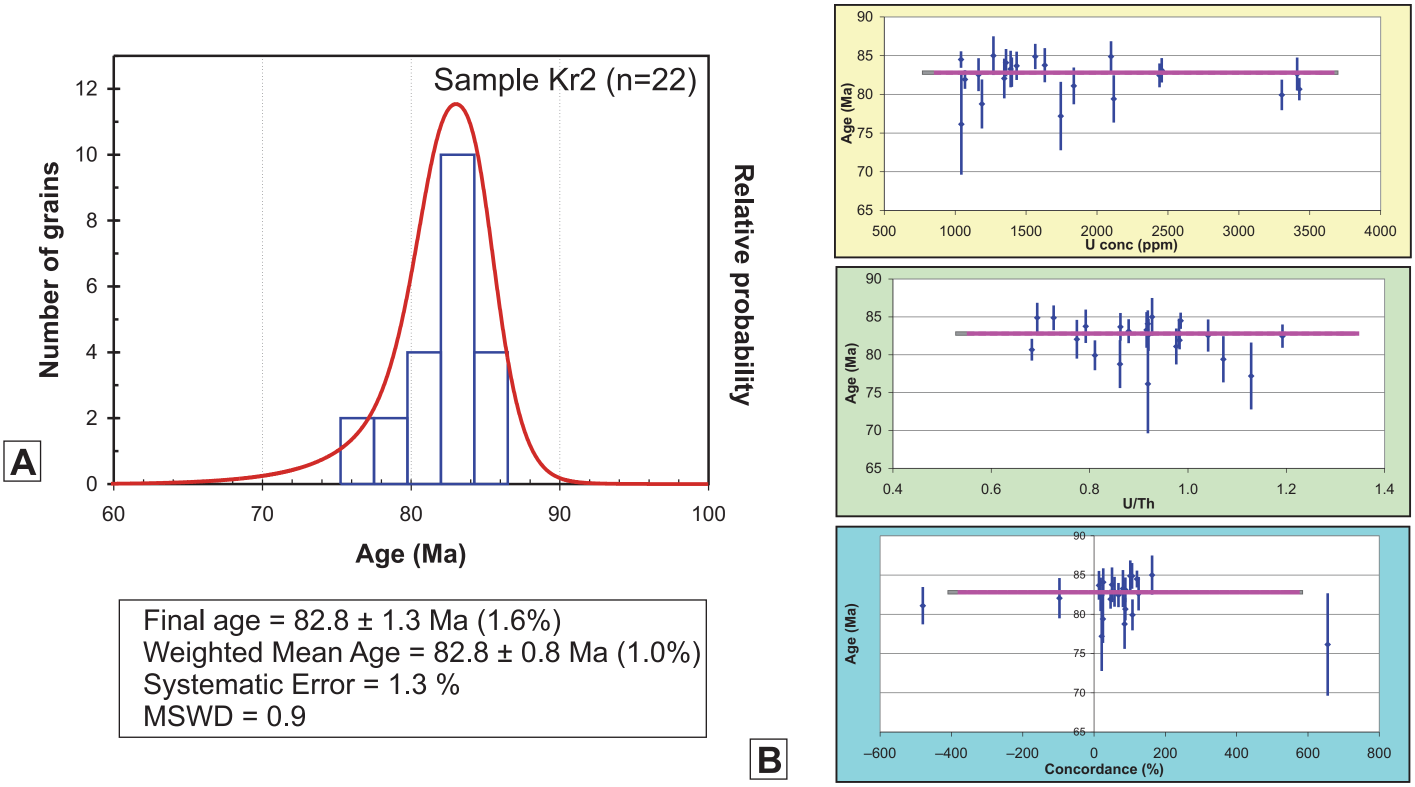


Figure 14. (A) U-Pb relative probability plot of 22 analyzed zircon crystals. MSWD—mean square of weighted deviates. (B) Ages are from the Age Pick program of G. Gehrels (University of Arizona). Blue error bars are at 1σ. Uncertainties for final age and weighted mean are at 2σ; conc.—concentration. Location of sample is in Figure 6.

Porcupine Hills Formation are tilted as a result of wedging of blind thrust sheets. These are the youngest preserved deposits that were horizontally shortened in the region. The youngest synorogenic units in the foreland are the Paleocene–Lower Eocene Fort Union and Wasatch Formations, preserved on the flanks of the Bearpaw Mountains and in the Missouri Breaks diatremes (Hearn, 1968, 1976; Hearn et al., 1964). Coeval deposits along the front of the thrust belt possibly accumulated in a wedge-top depozone, and were subsequently eroded. Erosion of the proximal foreland basin was a regional process. Nurkowski (1984) estimated that 900–1900 m of stratigraphic overburden was eroded from above near-surface coals in the Alberta Plains. As mentioned here, more recent thermal and forward kinematic modeling in southern Canada suggests ~3 km of postorogenic erosion in the proximal foredeep and 3–5 km of exhumation in the foothills (Hardebol et al., 2009).

Sandstone samples of the Fort Union, Porcupine Hills, and Wasatch Formations show an increase in the proportion of quartz relative to lower stratigraphic units (Fig. 12). Fort Union deposits are considerably sandier than underlying units, even at distal locations. The increase in the sandstone/shale ratio in this unit may be related to deeper exhumation of Belt Supergroup rocks in the hanging wall of the Lewis thrust system, which also could have increased the abundance of quartz and low-grade metamorphic clasts as byproducts of Belt quartzite and argillite erosion. The increased quartz content may also result from greater transport distance, insofar as these units were sampled at distal positions in the basin, or enhanced chemical weathering. The uppermost Wasatch Formation contains abundant volcanogenic sediment, as seen in multiple bentonite beds. Conglomerate clast counts from this unit indicate derivation from Belt Supergroup and volcanic rocks. Pebble conglomerates in the Wasatch Formation are dominated by clasts of Belt Supergroup quartzite (~28%) and fine-grained mafic igneous rocks (~28%). Shale, mudstone, and low-grade metamorphic pebbles constitute 17% of the total, followed closely by limestone clasts (16%). The remaining ~10% consists of similar proportions of fine-grained felsic igneous, granite, and schist clasts. Estimates of clast composition in Wasatch conglomerates by Hearn et al. (1964) provided values of 50%–80% argillite and quartzite of Belt Supergroup, 20%–40% Late Cretaceous–Paleocene fine-grained and porphyritic volcanic rocks, 1%–5% Paleozoic rocks, and 1% chert and conglomerate.

Synorogenic sedimentation in the foreland continued until the Early Eocene. Extensional faulting along the fold-thrust belt initiated soon after, during Middle Eocene time (Constenius, 1996). The development of extensional basins in the thrust belt overlapped in time with low-angle detachment faulting and development of metamorphic core complexes in hinterland regions (Doughty and Price, 2000; Johnson, 2006).

DISCUSSION

Controls on Thrust Belt Regional Geometry

The geometry, structure, and stratigraphic architecture of the Belt-Purcell basin controlled the architecture of the Cordilleran thrust belt of north-central Montana and southern Canada (Price and Sears, 2000). North of the Lewis and Clark line, the Belt-Purcell deposits form a thick, rheologically strong prism of quartz-rich metasedimentary rocks and mafic sills (Sears, 2007) with relatively minor internal shortening (Figs. 5 and 7). As suggested for other major

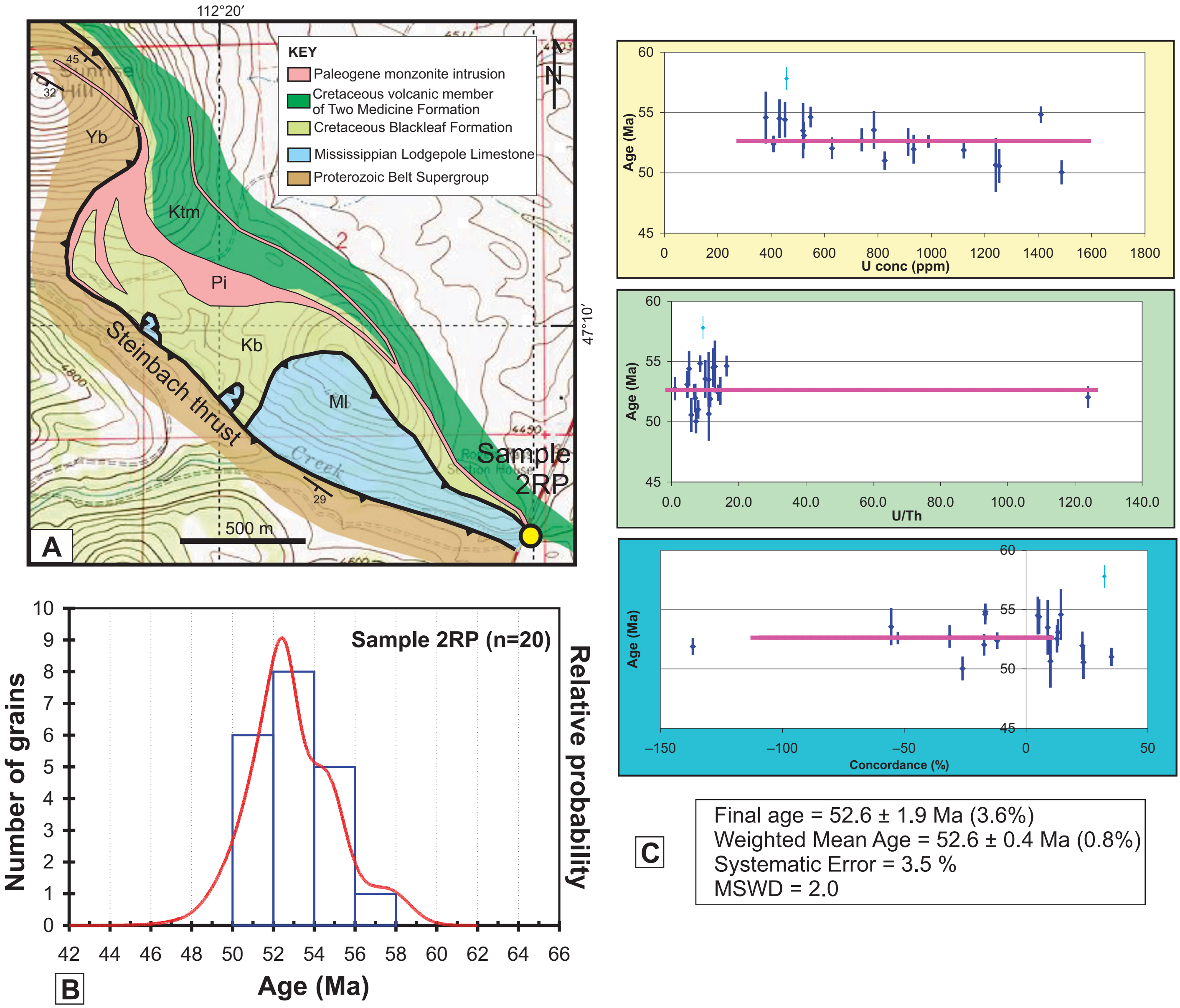


Figure 15. (A) Simplified map showing structural relations between the Steinbach thrust and the monzonite intrusion (after Harlan et al., 2005; based on R.G. Schmidt, personal commun.) in the Roger Pass area. In the northwest corner of the map, the intrusion cuts across the trace of the fault as a high-angle intrusion. In the hanging wall of the Steinbach thrust, deposits of the Belt Supergroup show thermal alteration. A complete discussion of the structural relation between the intrusion and the thrust was provided in Harlan et al. (2005). Also shown is location of sample 2RP. Approximate map location is in Figure 2. (B) U-Pb relative probability plot of 20 analyzed zircon crystals. (C) Ages are from the Age Pick program of G. Gehrels (University of Arizona). Blue error bars are at 1σ (light blue sample excluded). Uncertainties for final age and weighted mean are at 2σ. Previous K-Ar reported age was 59.6 ± 1.6 Ma (Schmidt, 1978; Whipple et al., 1987); conc.—concentration; MSWD—mean square of weighted deviates.

thrust sheets in the Cordilleran thrust belt (Mitra, 1997), the relative lack of internal shortening could be a result of this rheological strength. Sedimentary basin taper, or the predeformation basement dip, may be another key control in the structural style (Boyer, 1995). A hinterland increase in basement dip is characteristic of most basins deformed by thrust belts (Boyer, 1995); hence, trailing parts of thrust belts attain critical taper with relatively minor internal deformation, provided the rocks involved are rheologically strong (DeCelles and Mitra, 1995; Mitra, 1997; Yonkee, 2005). In northwestern Montana, this may be particularly true; deposits of the Belt Supergroup thicken markedly toward the west, probably the result of the transition from shelf to slope and basin along the preorogenic rifted basin margin. Flexure beneath the Paleozoic miogeocline and early foredeep depozone would have further increased basement dip toward the hinterland. Calculations of the depth to basement from seismic profiles and wells along our cross section support an increase of ~0.7° in the dip of the basal décollement (β) over a relatively short distance, from ~3.3° beneath the frontal part of the thrust belt to ~4° beneath the Lewis thrust system. In reality, the basement dip along any regional section varies with time once an orogenic system is established. This owes to the complex interaction of sedimentary and tectonic loads, the propagation toward the foreland of the fold-thrust belt and the flexural wave of the foreland basin, and possible dynamic subsidence. Thus, a westward increase in β was probably a major control in the low magnitude of internal shortening west of the leading edge of the Lewis thrust system.

Toward the foreland, the decrease in original dip of the basement-cover interface and the relatively thin prism of lithologically and rheologically variable Paleozoic and Mesozoic strata available to construct the thrust belt likely required the formation of the tightly imbricated stack of thrust sheets in the Sawtooth Range in order to maintain critical taper. In other words, high initial taper and great internal strength characterized the Lewis thrust system, whereas lower initial taper and generally lower strength but greater rheological heterogeneity (i.e., layering) characterized the Sawtooth Range system. The value of taper necessary for continued thrust belt forward propagation was nevertheless attained, as the thrust belt remained active until Eocene time.

Inherited structures of the Proterozoic rifted margin also controlled the structure of the thrust belt hinterland (Harrison et al., 1974; Winston, 1986). The Snowshoe and Pinkham thrusts have been interpreted as inverted Proterozoic normal faults (Sears, 2007). A question that arises with these interpretations, however, is the reason for the selective reactivation of only the upper segments of these structures, without basement involvement. Several thrust segments with anomalous orientations seem to represent inversion of synextensional transform faults (Price and Sears, 2000; Sears, 2007); a good example is the northeast-striking segment of the Moyie thrust in Canada, aligned with the crustal discontinuity of the Vulcan suture zone (Price and Sears, 2000).

A major difference between this sector of the Cordilleran thrust belt and areas of the Sevier belt farther south is in the timing of slip on the regional-scale, far-traveled Proterozoic quartzite-bearing thrust sheets. In northern Utah and Idaho, the Paris-Willard-Meade thrust system carries a thick succession of Proterozoic quartzite and overlying Paleozoic shelfal strata, with more than 50 km of slip (Yonkee et al., 1997). In central and southern Utah, the Canyon Range and Sheeprock thrusts and their lateral equivalents carry thick Proterozoic quartzite sections more than 100 km to the east. Throughout the Sevier thrust belt, these major thrust sheets were actively emplaced during Early Cretaceous (and possibly latest Jurassic) time (e.g., Yonkee, 1992; DeCelles, 2004). In contrast, the Lewis thrust system in northwestern Montana was mainly active during Campanian–Paleocene time. Whereas older thrust faults in northwestern Montana and adjacent British Columbia involved Proterozoic rocks, it was not until the Lewis thrust system became active that Proterozoic rocks were transported >100 km eastward.

Another key difference between this part of the Cordillera and other more intensively studied regions is the history of arc magmatism; in particular, the Cretaceous magmatic arc was situated directly west and southwest of the northwestern Montana thrust belt. High-volume magmatism in the region took place mainly shortly after the emplacement of the Lewis thrust system (Gaschnig et al., 2010), and may have been fueled by melt-fertile lower crust and lithosphere that was underthrust westward while the Lewis thrust system was active in the upper crust.

Relations with the Foreland Basin System

The foreland basin system records major tectonic events in the retroarc region. Jurassic strata of the Ellis Group and Morrison Formation in Montana were deposited in the distal part of an early foreland basin (possibly in a slowly subsiding back-bulge depozone) during collision of the Intermontane terrane (Fuentes et al., 2009, 2011) (Fig. 11). The regional unconformity at the top of the Morrison Formation originated in response to cratonward migration of a flexural forebulge, eustatic sea-level fall, and possible decreased dynamic subsidence. The Lower Cretaceous Kootanai Formation is the oldest unit of clearly synorogenic sedimentary strata that consistently thickens to the west, as expected for foredeep deposits. Provenance information indicates that sediment was derived from deformed miogeoclinal strata, volcanic sources in the magmatic arc, and the Intermontane belt. At that time, the tectonic subsidence curve shows the onset of a convex-upward pattern, typical of a foredeep depozone. All these elements suggest the propagation of the thrust belt toward the east and lateral migration of the flexural wave.

The Albian–lower Cenomanian Blackleaf Formation shows an abrupt increase in the proportion of low-grade metamorphic lithic fragments, suggesting involvement of Belt Supergroup rocks in thrust sheets in the Cordilleran hinterland. Candidate thrusts include the Hall Lake, St. Mary, and Moyie thrust system. Alternative or additional sediment sources include the transpressive structures in the Lewis and Clark line and western structures of the Helena salient region.

Deposition of the thick offshore facies of the Cenomanian–Santonian Marias River Shale, and similar deposits (such as the lower part of the Mancos Shale) are conspicuous along the Western Interior basin, and reflect high eustatic sea-level conditions (Miall et al., 2008).

The late Santonian–early Campanian interval was marked by a strongly regressive system, with rapid, relatively-coarse grained, clastic sedimentation. Thrust systems possibly active during this period include the Moyie, Snowshoe, Libby, Pinkham, Whitefish, and Wigwam. Concurrently, the Lewis and Clark strike-slip system was active, and foreland basin deposits north and south of it show differences in thickness and facies (Wallace et al., 1990; Sears and Hendrix, 2004).

Major slip on the Lewis thrust system commenced during deposition of the late Campanian Bearpaw-Horsethief Formation. An increase in low-grade metamorphic grains in sandstones of the lower part of the latest Campanian–Maastrichtian St. Mary River Formation suggests exhumation of Belt rocks along the Lewis system. Detrital zircons from the St. Mary River Formation indicate predepositional and syndepositional volcanic sources, as well as input from Paleozoic and Belt Supergroup strata. Both the St. Mary River and Willow Creek Formations indicate high sedimentation rates proximal to the advancing thrust front. The paradoxical absence of coarse-grained conglomerate in these formations may be explained by: (1) most material initially eroded from the hanging wall of the Lewis thrust system was fine-grained or unconsolidated foreland basin deposits, not prone to generating gravel; (2) coarse-grained alluvial fans were restricted to the most proximal areas in front of the Lewis thrust system, where they were vulnerable to erosion during continuing displacement on the Lewis and underlying thrusts in the Sawtooth Range duplex; and (3) coarse-grained sediment accumulation in the foreland basin was highly localized along the front of the Cordilleran thrust belt and was partly controlled by local across-strike structural discontinuities (Lawton et al., 1994).

The final record of synorogenic sedimentation in the foreland extended from Paleocene until the Early Eocene, during the youngest movement of the Lewis thrust system and deformation along the Sawtooth Range, and is recorded in erosional remnants of the Fort Union and Wasatch Formations in the distal foreland. Any coarse-grained, more proximal deposits were removed by uplift and erosion during postglacial isostatic rebound, erosion-driven isostatic rebound of the thrust belt (Sears, 2001), or exhumation in response to generalized extension since the Middle Eocene (Hardebol et al., 2009). Regional extensional faulting along the thrust belt commenced during Middle Eocene time (Constenius, 1996) and brought to a close the preceding ~110 m.y. episode of growth in the Cordilleran orogenic belt.

CONCLUSIONS

The Cordilleran retroarc at the latitude of northwestern Montana started its evolution during the Middle Jurassic. Although the early structures of the thrust belt have been removed by erosion, the foreland basin system records the initial stages of deformation in regions that are now located in the orogenic hinterland.

The westernmost preserved structures, including the Moyie, Snowshoe, Libby, and Pinkham thrust systems, deformed a thick (>15 km) package of Proterozoic Belt Supergroup deposits, possibly between the late-Early Cretaceous and the Campanian. These structures accommodate moderate amounts of shortening. The dominant structure along this segment of the thrust belt is the Lewis thrust system, which juxtaposes a relatively undeformed, rheologically strong package of Precambrian and Paleozoic quartzite, carbonate, and argillite on top of deformed Paleozoic and Mesozoic rocks in its footwall. To the east, the basal detachment climbs from Proterozoic into Cambrian levels to deform Paleozoic shelf strata via numerous, closely spaced thrusts of the Sawtooth Range. The frontal, eastern part of the thrust belt is characterized by highly deformed Mesozoic strata, with blind structures involving Paleozoic rocks at depth, as shown by seismic profiles. Shortening on the Lewis thrust system, and thrusts in the Sawtooth Range and foothills occurred roughly between mid-Campanian and Early Eocene time (ca. 75–52 Ma).

The enormous (>15 km structural thickness), relatively undeformed structure carried by the Lewis thrust system seems to be a result of two factors: the great rheological strength of the Belt Supergroup rocks in its hanging wall, and the relatively high angle dip of the basement-cover interface prior to Cordilleran shortening (Boyer, 1995), which helped this part of the thrust belt to attain critical taper values without much internal deformation.

Total shortening calculated from an ~145-km-long balanced cross section, from the undeformed foreland to the Mission Range, is ~135 km. This value is similar to shortening amounts for equivalent positions in the thrust belt documented farther north to the Canadian border (Bally, 1984; van der Velden and Cook, 1994). It is possible that the magnitude of shortening substantially decreases southward in Montana, as suggested by Price and Sears (2000) and Sears (2001, 2007).

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