# Subsidence across the Antler foreland of Montana and Idaho: Tectonic versus eustatic effects

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**Abstract** Devonian and Mississippian sedimentary rocks of western Montana and east-central Idaho were deposited on a cratonic platform that faced a deep basin to the west. The deep basin in Idaho was a northern extension of the Antler foredeep and formed as a flexural response to loading of the ancient North American continental margin by an inferred arc and thrust belt complex. Subsidence analyses of the Devonian–Mississippian strata indicate episodic subsidence events in the proximal foredeep and adjacent cratonic platform, an area approximately 800 km (500 mi) wide (palinspastic). Isopach maps for this sequence illustrate that many depocenters and paleohighs were geographically coincident across the foreland through time. The Devonian–Mississippian foreland structures had cross-sectional wavelengths of 50–200 km (30–120 mi) and amplitudes of about 50–350 m (160–1,150 ft). Some of these structures were tectonically inverted (i.e., paleohighs became depocenters and vice versa) several times during the 50–60 m.y. represented by this stratigraphic sequence. Many of these generally east-west-trending paleostructures were oriented at high angles to the north-south-trending axis of the Antler foredeep and the inferred strike of the Antler orogenic belt. These foreland structures coincide geographically with structural trends produced during Proterozoic extension, suggesting that the Proterozoic faults were reactivated during Antler convergence. The isopach maps also show progressive southeastward migration of Antler foredeep depocenters from Late Devonian to Early Pennsylvanian time. The southeastward migration of the foredeep depocenter suggests that the maximum thrust load moved progressively southeastward from Late Devonian to Early Pennsylvanian time. The complex patterns of subsidence across the Montana–Idaho foreland do not fit into simple flexural models for vertical loading of unbroken elastic plates. Instead, differential subsidence of the foreland may be related to several mechanisms: (1) flexure of mechanically independent, fault-bounded segments of the foreland produced by areally limited thrust loads (subregional vertical loading); (2) transmission of compressive in-plane stresses through the foreland lithosphere (regional horizontal loading) that may have reactivated Proterozoic fault systems; and (3) waxing and waning of in-plane compressive stresses resulting from the episodic nature of Antler convergence. Results from this study suggest that, in settings where the foreland lithosphere is broken by ancient fault systems, the foreland may exhibit complex patterns of differential subsidence that probably reflect a composite response to both vertical and horizontal loads. Also, the simultaneous pulses of subsidence documented across large parts of the Antler foreland suggest that it may be possible to date episodes of convergence along ancient continental margins, even when the ancient thrust belt complex is poorly preserved.

Foreland basins form largely as a flexural response to loading of continental lithosphere by thrust sheets (Beaumont, 1981; Jordan, 1981). Numerical models have been developed that relate the scale of the thrust load and rheologic properties of the loaded lithosphere to the stratigraphy and large-scale structures of the foreland area (Quinlan and Beaumont, 1984; Schedl and Wiltschko, 1984; Stockmal et al., 1986; Beaumont et al., 1988). These theoretical models of flexural behavior assume that the foreland lithosphere is not broken by lithosphere-scale faults that might prevent flexure from being transmitted uniformly across the foreland [cf. Stockmal and Beaumont (1987) and Royden et al. (1987)].

In addition to the flexural response of the foreland area, convergence along a continental margin produces horizontal in-plane compressive stresses that can be transmitted through foreland lithosphere. These stresses may affect lithosphere many hundreds of kilometers inboard of the actual thrust load and proximal foredeep (Lambeck et al., 1984; Cloetingh, 1988). As a result, horizontal in-plane stresses might enhance deflections of the crust in distal foreland or cratonic areas and therefore also affect sedimentation far from the zone of active plate margin convergence.

In this article we discuss the subsidence history and stratigraphic relationships of Middle Devonian through uppermost Mississippian sedimentary rocks in Montana and Idaho. This stratigraphic interval was deposited during a critical stage in the evolution of the North American Cordillera, namely the Antler orogeny. The effects of the Antler

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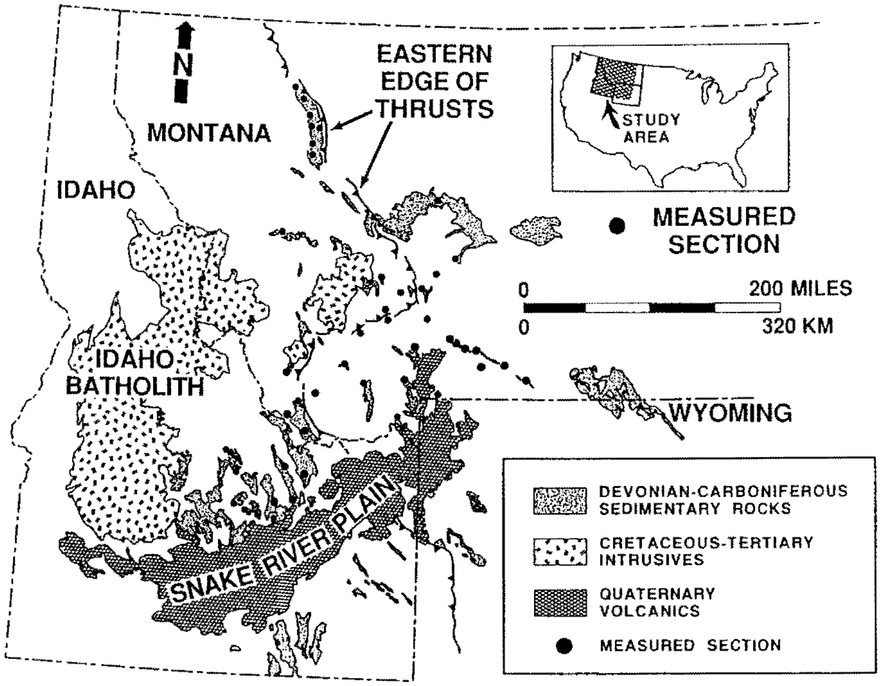



**Figure 1.** Location of study area.

orogeny on the foreland areas of Montana and Idaho are poorly understood. Therefore this study is important for documenting the stratigraphic record and subsidence history of the foreland and adjacent cratonic platform during Antler convergence. Results from our subsidence analyses allow interpretation of the relative effects of tectonic subsidence on the development of the foreland and cratonic stratigraphic sequences. This study also illustrates how subsidence of cratonic platforms can be affected by convergence events that occurred hundreds of kilometers outboard of the craton. Finally, this study suggests that, in convergent orogens where the fold and thrust belt is poorly preserved, convergence episodes can be dated indirectly from subsidence histories and sedimentologic analysis of *distal* foreland stratigraphic sequences.

## Structural and stratigraphic setting

**Regional structural history** Devonian through Mississippian sedimentary rocks are exposed throughout central Montana and east-central Idaho (fig. 1). In east-central Idaho and southwestern Montana most exposures of Devonian and Mississippian rocks are allochthonous, with local windows of possible parautochthonous rocks. Eastward transport of these rocks occurred along complex thrust and tear fault systems during Pennsylvanian (?) to early Tertiary time (Sevier and Laramide orogenies) (Skipp and Hall, 1975; Skipp et al., 1979; Ruppel and Lopez, 1984; Perry et al., 1989); evidence for any pre-Pennsylvanian shortening is equivocal (Dover, 1980). East of the leading edge of this fold and thrust belt, the amount of shortening is probably not significant enough to seriously affect palinspastic reconstructions. However, estimates for the total amount of tectonic shortening in the fold and thrust belt are highly variable and range from tens to hundreds of kilometers (Skipp and Hait, 1977; Nilsen, 1977; Skipp et al., 1979; Dover, 1980; Ruppel et al., 1981; Schmidt and Hendrix, 1981; Woodward, 1981; Skipp, 1988).

Late Tertiary extension produced the present north- or northwest-trending, block-faulted mountain ranges across the study area (Pardee, 1950; Reynolds, 1979; DuBois, 1983). The amount of extension progressively decreases eastward to the eastern edge of the study area. Movement along some of these high-angle faults still occurs, especially in east-central Idaho.

**Regional Devonian and Mississippian stratigraphic relationships** Devonian deposits in the study area unconformably overlie much older rocks, from Precambrian siliciclastics to lower Paleozoic siliciclastics and carbonates (Sloss and Moritz, 1951; Scholten, 1957, 1960; Scholten and Hait, 1962; Sandberg, 1961; Churkin, 1962; Loucks, 1977; Ruppel, 1986). Pre-Devonian subcrop maps (Sandberg and Mapel, 1967; Baars, 1972; Peterson, 1986) and field observations (Scholten, 1957, 1960; Churkin, 1962; Ruppel, 1986) show that on a regional scale the oldest subcrops of Precambrian and Cambrian sedimentary rocks occur near the present Montana–Idaho border, just north of the Snake River plain (fig. 2). Progressively younger lower Paleozoic sedimentary

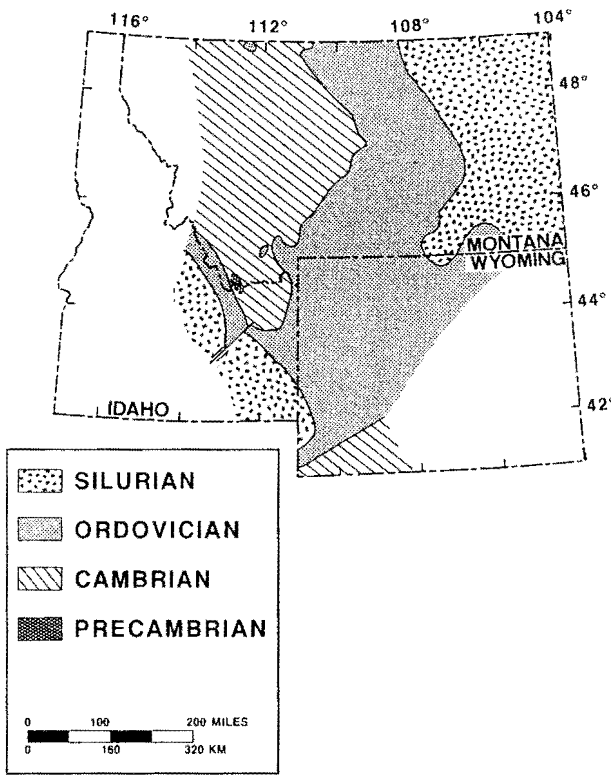



**Figure 2.** Pre-Devonian paleogeologic map. Modified from Sandberg and Mapel (1967) and Baars (1972) using data from Churkin (1962) and Scholten (1957, 1960) and unpublished field data. Note that the oldest rocks beneath Devonian deposits are located in southwestern Montana near the present Montana–Idaho border. Lower Paleozoic subcrop patterns are highly generalized across the entire area.

rocks underlie Devonian deposits to the east and west of this subcrop trend.

The oldest Devonian deposits in the study area occur in large [up to 90 m (300 ft) deep and 650 m (2,100 ft) wide] paleovalleys (or channels) that are found in western Montana and east-central Idaho (fig. 3). These valleys are filled with Lower to Middle Devonian strata that are included with the Jefferson Formation, named separately (e.g., Beartooth Butte Formation), or not given formal stratigraphic names (Churkin, 1962; Scholten and Hait, 1962; Sandberg and Mapel, 1967; Mapel and Sandberg, 1968; Hoggan, 1981). Upper Devonian deposits consist of cyclic platform carbonates of the Jefferson Formation and shallow-water carbonate, siliciclastic, and evaporite deposits of the unconformably overlying Three Forks Formation. Unconformities separate the three members within the Three Forks Formation (in ascending order): the Logan Gulch, Trident, and Sappington Members. Minor unconformities also occur locally within the members. The regional extent and duration of these unconformities and lithostratigraphic relationships within the Three Forks Formation are highly variable (fig. 3) (Sandberg and Poole, 1977; Sandberg et al., 1983) and are discussed in greater detail later. To the west of the study area, shales and siltstones of the Lower to Upper Devonian Milligen Formation [1,200+ m (3,900+ ft) thick] were deposited before and during deposition of the Middle–Upper Devonian units on the Montana platform and eastern side of the Antler foredeep.

Lower Mississippian carbonates and siliciclastics of the Madison Group and equivalent units unconformably overlie the Three Forks Formation across the study area (fig. 3). In Montana the Madison Group consists of the Lodgepole and Mission Canyon Formations. The Lodgepole Formation [150–300 m (490–980 ft) thick] has three conformable members (in ascending order): Cottonwood Canyon Member, Paine Member, and Woodhurst Member. The Cottonwood Canyon Member is a thin sequence [generally <5 m (<16 ft) thick across most of the study area] of black shale, siltstone, sandstone, and dolostone that has been interpreted as a shallow marine to marginal marine deposit (Gutschick et al., 1976; Sandberg and Klapper, 1976). The Cottonwood Canyon Member also has been interpreted as a condensed sequence based on the presence of glauconite, phosphatic nodules, and abundant conodonts and fish remains (Sandberg and Klapper, 1976; Sandberg et al., 1983). The Paine Member [50–90 m (160–300 ft) thick] typically consists of thin-bedded, shaly limestone with isolated Waulsortian-type mud mounds. The Woodhurst Member [100–170 m (330–560 ft) thick] consists of cyclic deep ramp deposits, including thin-bedded shaly limestone through coarse grainstone facies. Overall, the Lodgepole Formation represents a shallowing-upward onlap sequence or transgressive systems tract that contains superimposed smaller-scale sequences [in the sense of Van Wagoner et al. (1988)]. Basinal stratigraphic equivalents of the Lodgepole Formation include (1) the lower member of the McGowan Creek Formation, a thick [100–800+ m (330–2,600+ ft) thick] shale and siltstone flysch sequence that was deposited in the deeper axial parts of the Antler foredeep in east-central Idaho, and (2) the lower part of the Middle Canyon Formation [50–200 m (160–660 ft) thick], a dark cherty limestone sequence that was deposited in outer ramp and slope environments.

The Mission Canyon Formation [200–400 m (660–1,300 ft) thick] and its stratigraphic equivalents conformably overlie the Lodgepole and Middle Canyon (part) Formations (fig. 3). The Mission Canyon Formation is a highstand systems tract [in the sense of Van Wagoner et al. (1988)] that occurs throughout central and southwestern Montana and consists of cyclic, shallow subtidal to peritidal platform facies (Reid and Dorobek, 1989, 1991). These cyclic facies grade westward into a thick but relatively narrow facies tract of skeletal-ooid grainstone and packstone, which in turn grades westward into equivalent slope and basinal facies of the Middle Canyon and upper member of the McGowan Creek Formations (fig. 3) (Huh, 1967, 1968; Sandberg, 1975; Skipp et al., 1979; Gutschick et al., 1980). The grainstone-packstone facies tract of the Mission Canyon Formation represents a series of

A

B

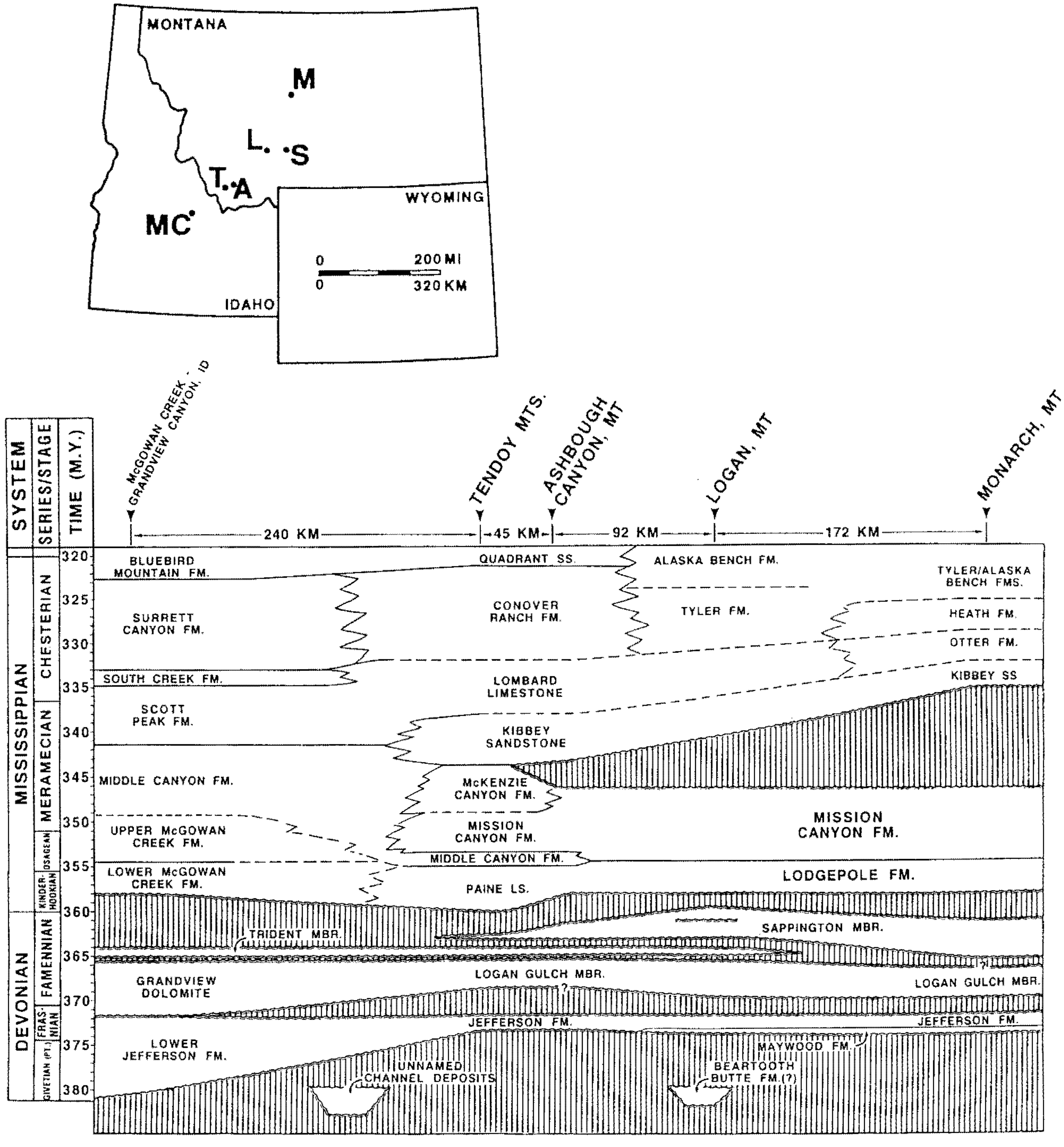



**Figure 3.** (A) Location map for selected localities used to construct chronostratigraphic chart shown in part B. M, Monarch; S, Sacajawea Peak; L, Logan; A, Ashbough Canyon; T, Tendoy Mountains; MC, composite section from McGowan Creek and Grandview Canyon. These selected localities occur along a platform to basin transect. Measured section locations on this base map also were chosen for subsidence analyses; subsidence curves for each locality are shown in fig. 7. (B) Chronostratigraphic chart for Upper Devonian to Upper Mississippian strata across the study area. Distances between localities used to construct the chronostratigraphic chart are palinspastic [palinspastic base from Peterson (1986)]; distances on the base map shown in part A are nonpalinspastic distances. Unconformities indicated by vertically ruled areas. Logan Gulch, Trident, and Sappington Members make up the Three Forks Formation, which is discussed in the text but not labeled in this figure. Note highly variable durations of many unconformities on a regional scale, especially many of the Late Devonian unconformities. Lateral contacts between formations often are dashed because stratigraphic interfingering between many units has not been observed in this structurally complex area.

overlapping carbonate-sand shoals that developed along the western margin of the Early Mississippian platform (Rose, 1976; Sandberg et al., 1983). Eastern exposures of the Middle Canyon Formation consist of cherty limestones that grade westward into silty limestones and calcareous siltstones of the upper McGowan Creek and Middle Canyon Formations (fig. 3). Paleocurrent data from Lower Mississippian strata in east-central Idaho indicate that siliciclastic turbidites from the western side of the foredeep were derived from Antler highlands, whereas carbonate turbidites from the eastern side of the basin were derived from the Mission Canyon platform (Nilsen, 1977; Reid, 1991).

A regional unconformity occurs on top of the Mission Canyon Formation throughout Montana. This unconformity represents 9–14 m.y. of subaerial exposure (Sando, 1976, 1988; Skipp et al., 1979; Gutschick et al., 1980; Sandberg et al., 1983), during which the Mission Canyon platform underwent extensive karstification (Middleton, 1961; Roberts, 1966; Sando, 1974, 1988). However, this regional unconformity does not extend basinward of the platform margin carbonate-sand shoal facies belt in the Mission Canyon Formation and therefore represents a type 2 sequence boundary [in the sense of Van Wagoner et al. (1988)]. Peritidal facies of the McKenzie Canyon Formation (fig. 3) and time-equivalent deeper water facies of the upper Middle Canyon Formation were deposited during subaerial exposure of the Mission Canyon platform in middle to upper Meramecian time (Sando et al., 1985). Shallow-water facies of the McKenzie Canyon Formation were deposited peripheral to the earlier margin of the Mission Canyon platform (Sando et al., 1985; Reid and Dorobek, 1989) and may represent a platform margin carbonate wedge [in the sense of Sarg (1988)].

Uppermost Mississippian (upper Meramecian to Chesterian) sedimentary rocks overlie the Madison Group and Middle Canyon Formation across the study area (fig. 3). These units consist of the Big Snowy Group [shallow-water clastics and carbonates; 0–350 m (0–1,200 ft) thick] in central and southwestern Montana; the Snowcrest Range group [shallow-water clastics and carbonates; ~230 m (~750 ft) thick; Sando et al., 1985] in southwestern Montana; the White Knob Group [shallow to deep ramp carbonates and clastics and basinal facies; 1,000–1,700+ m (3,300–5,600+ ft) thick] in east-central Idaho (Huh, 1967, 1968); and the Copper Basin Formation [conglomerate, sandstone, argillite, and minor limestone; 2,000–3,000+ m (6,600–9,800+ ft) thick] in the westernmost part of the study area (Paull et al., 1972; Skipp et al., 1979). Thicknesses and lithologic data from these Upper Mississippian stratigraphic units were taken from previous studies and used in our subsidence analyses.

## Devonian–Mississippian paleogeography and tectonic setting

**Late Proterozoic through Mississippian tectonic history** Before Middle Devonian time, a passive continental margin existed across Montana and east-central Idaho. This passive margin was initiated during late Proterozoic–Early Cambrian rifting (600–550 Ma) (Stewart and Suczek, 1977; Armin and Mayer, 1983; Bond et al., 1983; Bond and Kominz, 1984) and persisted until Early to Middle (?) Devonian time. However, by Late Devonian–Early Mississippian time an inferred volcanic arc collided with the western margin of North America. This collision produced a foredeep basin that was superimposed on the underlying passive margin sequence. In general, the eastern side of this foredeep was located near the hinge zone of the antecedent lower Paleozoic passive margin; the western and deeper part of the foredeep formed above outer shelf to slope facies of the lower Paleozoic passive margin. This interval of Late Devonian–Early Mississippian convergence is known regionally as the Antler orogeny (Roberts and Thomasson, 1964; Burchfiel and Davis, 1972; Poole, 1974; Dickinson, 1977; Speed, 1977; Dover, 1980; Speed and Sleep, 1982). Similar convergence events affected much of the western margin of North America during Late Devonian–Early Mississippian time (Gordey, 1988; Morrow and Geldsetzer, 1988; Oldow et al., 1989), although unequivocal evidence for the arc-continent collision is poorly preserved along much of the North American Cordillera. Geologic evidence for the Antler orogeny is especially cryptic in east-central Idaho (Nilsen, 1977; Dover, 1980). Evidence for an Antler orogenic highland in Idaho is based largely on the preserved stratigraphy of synorogenic siliciclastic sediments (Devonian Picabo and Milligen Formations and Mississippian McGowan Creek Formation, White Knob Group, and Copper Basin Formation) in the proximal foredeep (Sandberg, 1975; Sandberg et al., 1975; Skipp and Sandberg, 1975; Nilsen, 1977; Dover, 1980). In summary, Devonian and Mississippian sedimentary rocks of southwestern Montana and east-central Idaho were deposited when an earlier passive margin was transformed into a convergent margin.

**Initiation of the Antler foredeep in Idaho: Stratigraphic constraints** The time of initiation and paleogeographic location of the Antler foredeep in Idaho can be constrained by regional stratigraphic relationships. The Lower to Upper Devonian Milligen Formation is a thick [>1,200 m (>4,000 ft)] sequence of argillite and sandstone that largely correlates with carbonate facies of the Carey Dolomite and Jefferson Formation to the east (Sandberg et al., 1975; Johnson et al., 1985; Dorobek, 1987), although lateral facies transitions have been obscured by several deformation events. In Middle Devonian (Givetian) time most of Montana was subaerially exposed while outer ramp to slope carbonate facies, which comprise the lower part of the Jefferson Formation, were being deposited in east-central Idaho. An eastern source area has been suggested for sediments in the Milligen Formation (Sandberg et al., 1975), but a western source may be more likely, given that the Jefferson Formation in east-central Idaho is nearly all dolomite and contains little shale. These stratigraphic relationships and the results from our subsidence analyses (see fig. 7) suggest that subsidence in the foredeep began earlier, in the Early to Middle Devonian, than on the cratonic platform and may reflect the initial response of the foredeep to the encroaching Antler accretionary wedge.

By late Frasnian–early Famennian time a subregional unconformity developed between the top of the Milligen Formation (earliest Frasnian) and an overlying unnamed limestone unit of Famennian age in the Wood River area of east-central Idaho (Sandberg et al., 1975; Johnson et al., 1985). Basinal facies of the Milligen Formation must have been uplifted to produce this unconformity. In addition, sandstone and restricted peritidal facies of the "Grandview Dolomite" (uppermost part of the Jefferson Formation in the Lost River Range of east-central Idaho) were deposited to the east during the time represented by this unconformity. "Grandview Dolomite" facies apparently prograded from west to east (P. E. Isaacson, personal communication, 1986); sandstone lithofacies in the "Grandview Dolomite" thicken westward, and sandstone paleocurrent data indicate west to east transport directions (Dorobek, unpublished data, 1986). Finally, these Late Devonian events precede deposition of conglomerate and sandstone of the upper Famennian Picabo Formation. Picabo siliciclastics also probably were derived from local uplifted areas associated with the Antler orogeny (Skipp and Sandberg, 1975; Isaacson et al., 1983).

These stratigraphic relationships suggest that an uplifted sediment source area existed to the west of the cratonic platform, possibly as early as the Early Devonian but clearly by Late Devonian time (Isaacson et al., 1983). A western source terrane also is well documented for the various Mississippian siliciclastic units (McGowan Creek Formation, White Knob Group, and Copper Basin Formation) that filled the Antler foredeep (Sandberg, 1975; Nilsen, 1977; Dover, 1980; Sandberg et al., 1983). These regional stratigraphic relationships are important for interpreting the subsidence curves from the Antler basin and cratonic platform discussed later.

**Devonian–Mississippian differential subsidence and tectonic inversion across the Montana–Idaho foreland** The Late Devonian through Late Mississippian platform, which extended across Montana and eastern Idaho, had complex paleotopography. Published isopach maps and regional lithofacies patterns indicate that Devonian–Mississippian sedimentation in the study area was affected by a number of low-amplitude paleohighs and troughs that dissected the foreland area (fig. 4). Some foreland structures were tectonically active during this time as far east as the Williston basin (McCabe, 1954; Clement, 1986; Gerhard et al., 1987; LeFever et al., 1987). Many of these Devonian–Mississippian paleostructures were oriented at high angles to the strike of the inferred Antler orogenic belt and adjacent foredeep.

Comparison of palinspastically restored isopach maps for Precambrian (Belt Supergroup), Devonian, Lower to Middle Mississippian, and Upper Mississippian to Lower Pennsylvanian stratigraphic units shows that the general trend of paleotopographic features across Montana remained essentially the same from Middle Proterozoic to Early Pennsylvanian time (fig. 4), especially across central Montana. In addition, these topographic features underwent several episodes of tectonic inversion. Tectonic inversion [in the sense of Visser (1980) and Ziegler (1987a,b)] refers to unspecified tectonic processes that result in a reversal of subsidence or uplift in a particular region (e.g., trough areas are uplifted and/or paleohighs are inverted and become depocenters).

The Precambrian structural grain that was reactivated during Antler time probably formed during a middle Proterozoic extensional event that produced several depocenters across Montana and Idaho (fig. 4) (Peterson, 1986; Tonnsen, 1986). This system of depocenters, which includes the middle to late Proterozoic Belt basin, may have formed above oceanic or extremely thin continental crust in a passive margin (McMechan, 1981), intracratonic rift (Winston et al., 1984), or remnant back-arc basin (Hoffman, 1989) setting. The Central Montana trough is an east-west-trending arm of the Belt basin that extended across central Montana (fig. 4). By analogy with other extensional settings, the Central Montana trough probably was bounded by normal faults that formed during middle Proterozoic extension.

The Central Montana trough again became an elongate depocenter after the 600–550-Ma rifting event that initiated the lower Paleozoic passive margin. Proterozoic faults apparently were reactivated during the 600–550-Ma extension, as indicated by the geographic coincidence of Proterozoic and lower Paleozoic depocenters across central Montana. The Central Montana trough remained a depocenter until the Ordovician. The absence of Ordovician to Silurian strata across much of western and central Montana suggests that central Montana probably was emergent from Ordovician to Early Devonian time. During the Middle Devonian much of the former Central Montana trough area was a paleohigh called the Central Montana uplift. By the Late Devonian, subsidence of the Central Montana uplift accelerated (discussed later), allowing progressive onlap of the Frasnian Jefferson Formation (Dorobek and Smith, 1989; Dorobek, 1991) and some units of the Famennian Three Forks Formation during the Late Devonian.

By Early Mississippian time the central Montana region was completely covered by the Lodgepole Formation, but

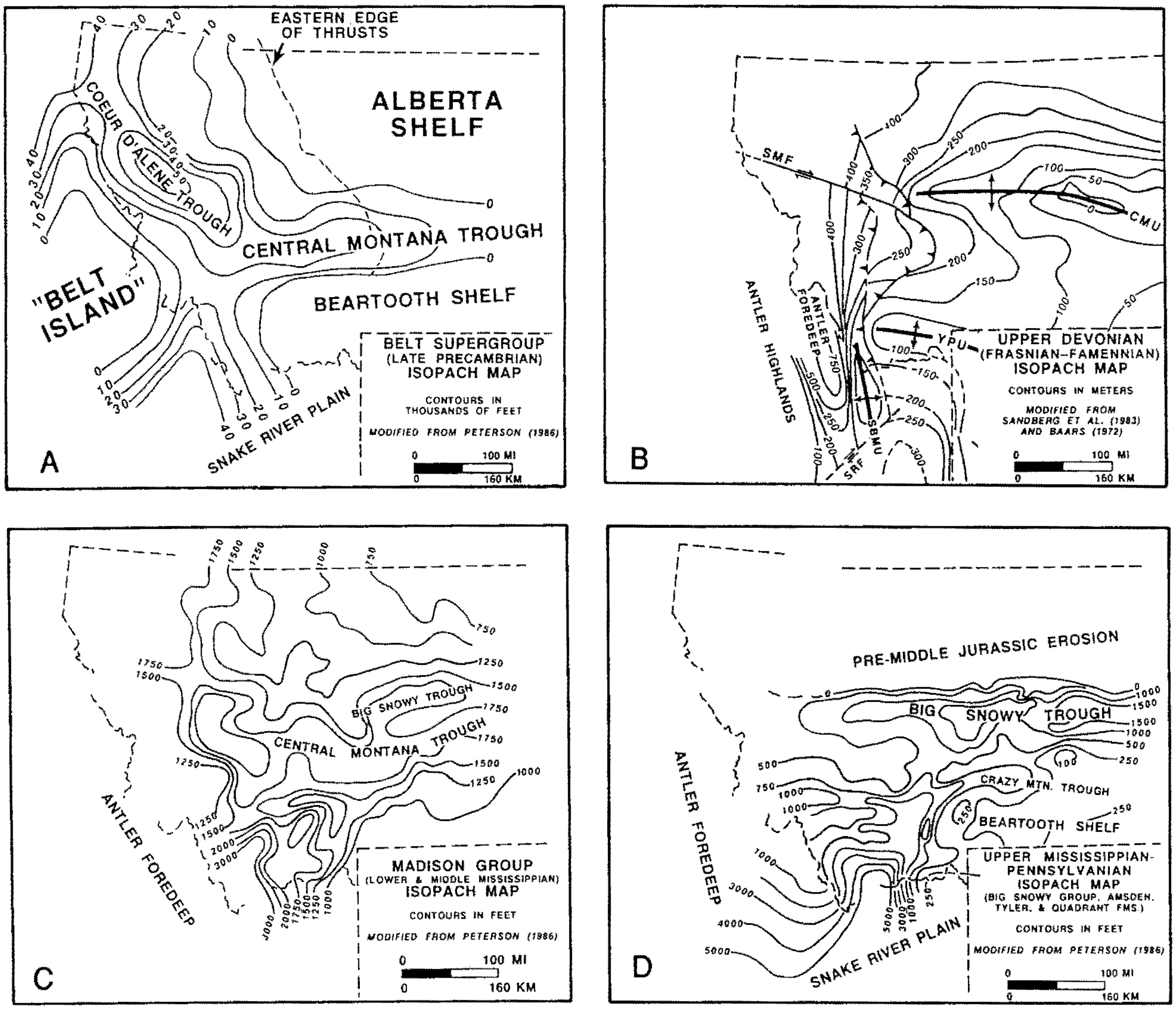



**Figure 4.** Isopach maps for specific time intervals; partly restored base. Major paleotopographic features are also indicated. Isopach thicknesses are reasonably accurate because Antler deformation never advanced far enough eastward during the Devonian–Mississippian to significantly deform and/or erode much of the sedimentary section that is contoured in these isopach maps. (A) Isopach map for Proterozoic Belt supergroup. Several depocenters make up the Belt basin; note especially the east-west-trending Central Montana trough. (B) Isopach map for Frasnian and lower Famennian deposits. SRF, Snake River fault zone; SMF, St. Mary's fault zone; CMU, Central Montana uplift; YPU, Yellowstone Park uplift; SBMU, Southern Beaverhead Mountains uplift. Thrust traces (with sawteeth) delineate major Laramide thrusts; these thrusts are shown here to be consistent with the original isopach map of Sandberg et al. (1983). Note that the Late Devonian Central Montana uplift coincided geographically with the Proterozoic Central Montana trough depocenter shown in part A. Also note the location of the Late Devonian foredeep depocenter. (C) Isopach map for Lower and Middle Mississippian Madison Group. Note that the Central Montana uplift, which was present across Montana in Late Devonian time (part B), underwent tectonic inversion and became the Central Montana trough in Early to Middle Mississippian time. (D) Isopach map for Upper Mississippian and Pennsylvanian rocks. Note that the foreland area in Montana still was dissected by numerous paleostructures that were oriented at high angles to the Antler foredeep axis. Also note that the Late Mississippian–Early Pennsylvanian foredeep depocenter was located much further to the southeast than the Late Devonian foredeep depocenter.

there still was enough topography to affect facies tracts within the Lodgepole Formation (Smith, 1977). Minor differential subsidence occurred in the central Montana region during deposition of the Mission Canyon Formation. Regional biostratigraphic and lithostratigraphic relationships also suggest that some deformation of the foreland platform occurred during Middle Mississippian time (Sando et al., 1975; Sando, 1988). During later Mississippian and Pennsylvanian time, central Montana again began to subside differentially, and a new depocenter, the Big Snowy trough, was superimposed over the former Central Montana uplift–Central Montana trough pair. The Big Snowy trough became the dominant depocenter across central Montana, allowing thick sequences of the Big Snowy Group and various Pennsylvanian units to accumulate (Craig, 1972; Mallory, 1972; Peterson, 1986).

The amplitude and wavelength of the paleostructures across the Montana–Idaho foreland can be obtained by constructing cross sections through the stratigraphic section for several 10–20-m.y. intervals (figs. 5 and 6). These cross sections of stratigraphic thickness show that the paleohighs and troughs across central Montana had (sediment-filled) cross-sectional amplitudes of about 50–350 m (160–1,150 ft) and wavelengths of 50–200 km (30–120 mi). These generally east-west-trending paleostructures also were oriented at high angles to the north-south-trending axis of the Antler foredeep and the inferred strike of the Antler orogenic belt.

The isopach maps also show a progressive southeastward migration of the Antler foredeep depocenters from Late Devonian to Early Pennsylvanian time (fig. 4). The greatest subsidence in foreland basins typically occurs adjacent to the maximum (i.e., thickest) thrust load (Jordan, 1981; Quinlan and Beaumont, 1984; Stockmal et al., 1986). Therefore southeastward migration of the foredeep depocenter suggests that the maximum thrust load also moved progressively southeastward from Late Devonian to Early Pennsylvanian time.

In summary, many of the paleotopographic elements that affected Devonian through Mississippian stratigraphy in the study area suggest reactivation of middle Proterozoic extensional faults during the Antler orogeny. Southeastward migration of Antler foredeep depocenters indicates southeastward migration of the zone of maximum thrust loading.

## Subsidence analyses across the Antler foreland

The subsidence history of the Antler foreland in Idaho and adjacent cratonic platform in Montana was highly variable, even over short distances. The restored stratigraphic cross sections in fig. 6 illustrate the effects of tectonic inversion across the study area. However, quantitative subsidence analyses (or backstripping) were done to understand the timing and relative rates of subsidence across the study area. Backstripping is performed by taking cumulative stratigraphic thicknesses at a particular geographic location and removing the effects of sediment lithification, sediment loading, and paleobathymetric changes during deposition of each stratigraphic interval (Sleep, 1971; Steckler and Watts, 1978; Bond and Kominz, 1984). This procedure generates a curve whose form reflects the tectonic or driving component of subsidence and eustatic sea-level change through time. The backstripping procedure used in this study has been described by Bond and Kominz (1984) and Bond et al. (1988). The model parameters used in this study's backstripping analyses have also been described elsewhere (Reid, 1991; Dorobek et al., 1991).

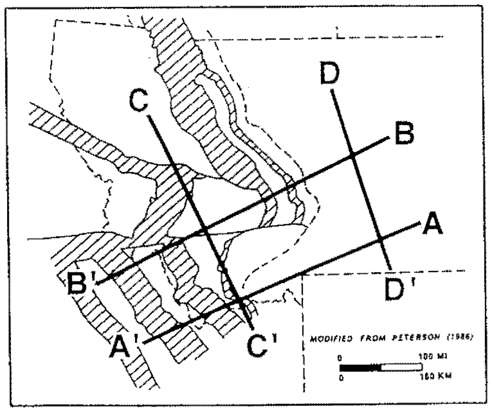



**Figure 5.** Palinspastic base map with locations of thickness cross sections shown in fig. 6. Hachured regions show amount of palinspastic restoration in overthrust belt. Cross sections A–A′ and B–B′ are dip sections; cross sections C–C′ and D–D′ trend approximately parallel to regional depositional strike. Palinspastic base from Peterson (1986).

**Subsidence curves from the Antler foreland and their interpretation** Six localities were selected for subsidence analyses (fig. 3A). These localities were selected because (1) they define a platform to basin transect, (2) the Devonian–Mississippian stratigraphy is well exposed at these localities, and (3) biostratigraphic boundaries are reasonably well constrained. Subsidence analyses were not attempted for the deepest part of the Antler foredeep because age determinations and stratigraphic thicknesses are poorly constrained.

Important similarities and differences exist between the platform and foredeep basin subsidence histories. Subsidence began at McGowan Creek–Grandview Canyon in the Middle Devonian (fig. 7A). However, farther west, subsidence began even earlier, as indicated by an Early Devonian age for the base of the Milligen Formation [inferred by Sandberg et al. (1975) based on early Middle Devonian conodonts that were found 300 m (980 ft) from the base of the Milligen Formation]. The Milligen Formation was deposited near western parts of the incipient Antler foredeep, suggest-

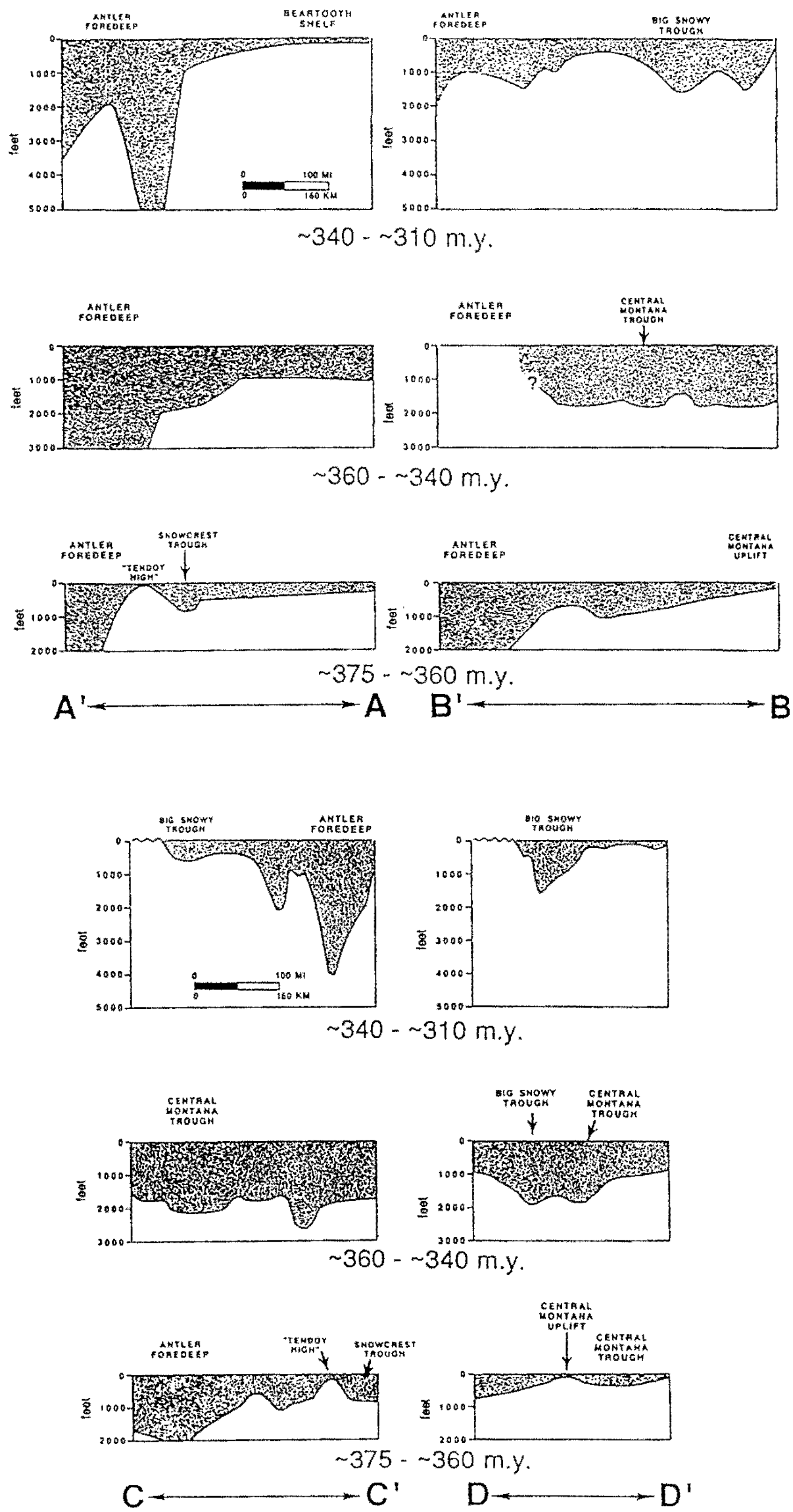



**Figure 6.** Cross sections illustrating compacted stratigraphic thicknesses (shown as shaded areas) across the study area for specific time intervals. 375–360 m.y. = Late Devonian to Devonian–Mississippian boundary; 360–340 m.y. = Devonian–Mississippian boundary to Middle Mississippian; 340–310 m.y. = Middle Mississippian to Early Pennsylvanian. Note tectonic inversion across the foreland area, which is especially notable on cross sections C–C′ and D–D′.

ing that subsidence in the foredeep began in the Early to Middle Devonian. Unfortunately, structural complications and poor exposure make quantitative subsidence analyses impossible in this part of the Antler foredeep. However, the presence of Lower to Middle Devonian flyschlike deposits suggest that similar pre-Antler tectonic events occurred in Idaho as in Nevada and Utah [cf. Hintze (1973) and Sandberg et al. (1983)].

In contrast to the Early to Middle Devonian onset of subsidence in the proximal foredeep, subsidence of the Montana platform apparently began in the Late Devonian (Frasnian) (figs. 7B–F). There are several possible explanations why subsidence apparently began at different times in the foredeep and adjacent platform. First, the earlier onset of subsidence in the proximal foredeep may reflect the initial response of the foredeep to the encroaching Antler accretionary wedge. The distal foreland of Montana simply may have been too far east to have been affected by flexure caused by the Antler accretionary wedge. This might explain the relatively high rate and magnitude of Early to Middle Devonian subsidence in the incipient Antler foredeep and the apparent lack of coeval subsidence on the platform.

Initial vertical loading by the Antler accretionary wedge also may have been located above continental lithosphere that had been thinned during late Proterozoic–early Paleozoic extension. It is possible that the region of initial vertical loading had relatively low flexural rigidity, thus preventing flexure of regions far inboard from the Antler accretionary prism. In this case the width of the zone of flexure would have been less than might be expected if thermally mature continental lithosphere were loaded. However, Antler convergence occurred approximately 150–200 m.y. after the 600–550-Ma rifting event. Continental lithosphere in Idaho and Montana should have been thermally mature by Antler time, making this second explanation unlikely.

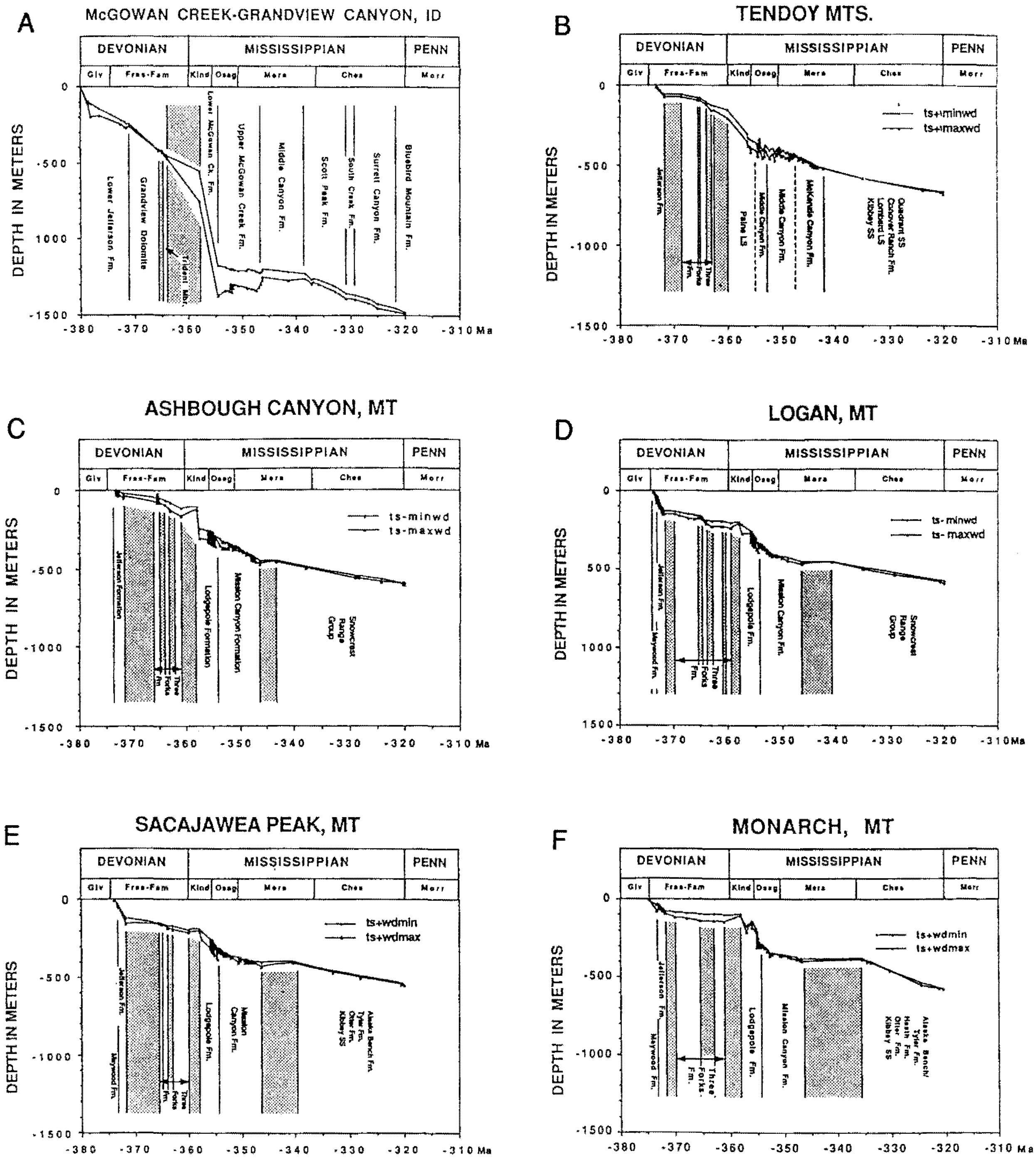
The apparent pulse of Frasnian–Famennian tectonic subsidence on the Montana platform also may have been caused by a rise in sea level. Bond and Kominz (1991) estimated from subsidence analyses of midcontinent strata in Iowa that the Frasnian–Famennian sea-level rise was ~100 m (~330 ft). The magnitude of Frasnian–Famennian subsidence also is of the order of 100 m (330 ft) for several platform locations in Montana (Tendoy Mountains, Ashbough Canyon, and Monarch). At the other platform locations (Logan and Sacajawea Peak), the Frasnian–Famennian subsidence pulse is close to the calculated upper limit of maximum sea-level rise (Bond and Kominz, 1991) if all reasonable sources of error are considered in the subsidence analyses from Iowa. Therefore the apparent increase in platform subsidence during Frasnian–Famennian time may largely record a Late Devonian sea-level rise and *not* the coupled tectonic response of the Antler accretionary wedge, foredeep, and distal foreland platform. However, although a sea-level rise may explain the apparent Late Devonian pulse of platform subsidence, the Frasnian–Famennian subsidence at McGowan Creek–Grandview Canyon (fig. 7A) is nearly twice the maximum estimated sea-level rise and probably reflects tectonically driven subsidence, at least in the proximal foredeep. Therefore the remaining question is whether or not the apparent pulse of Frasnian–Famennian subsidence in the distal Antler foreland reflects (1) a eustatic sea-level rise, (2) tectonic loading and mechanical coupling of the proximal foredeep and distal foreland, or (3) some combination of both mechanisms.

The remaining segments of higher subsidence on Mississippian parts of the subsidence curves from all localities greatly exceed the estimated sea level rise during Early Mississippian to Late Mississippian time (Bond and Kominz, 1991; Dorobek et al., 1991) and must be a response to tectonic loads. Subsidence curves for platform locations (Monarch, Sacajawea Peak, Logan, and Ashbough Canyon; figs. 7C–F) are similar in general form. Except for late Frasnian to middle Kinderhookian (early Mississippian) time, the position of Mississippian inflection points on the subsidence curves are remarkably consistent across the platform and from the most basinal locality (McGowan Creek–Grandview Canyon), suggesting that the proximal foredeep and distal foreland were mechanically coupled after the Famennian and responded to the same tectonic load(s) (fig. 7).

The tectonic loading produced a response over a vast region [over 800 km (500 mi) palinspastic distance]. However, this probably was not a purely flexural response across the entire study area because the wavelengths are too great to have been produced solely by flexure resulting from thrust loading along the western continental margin of North America. Many of the paleotopographic elements across the distal foreland also were oriented at high angles to the axis of the Antler foredeep, which would not be expected if these elements had been produced solely by flexure. The axes of the foreland structures would have been subparallel to the foredeep axis if they were entirely a flexural response to the vertical load of the Antler accretionary wedge.

Similarities in the subsidence curves across the Antler foreland provide some understanding of possible tectonic

**Figure 7.** Subsidence curves generated for selected localities shown in fig. 3A. Unconformities indicated by stippled intervals. Two subsidence curves are shown, one generated using corrections for minimum water-depth estimates for each stratigraphic unit ("ts-minwd"), and the other generated using maximum water depth estimates ("ts-maxwd"). Note that most facies were deposited in water depths less than 50 m (160 ft); therefore differences in curves resulting from water-depth corrections are negligible. See text for discussion on significance of inflection points on subsidence curves. Monarch, Sacajawea Peak, Logan, and Ashbough Canyon are platform localities. Tendoy Mountains section is a transitional area located near the transition from shallow platform to deeper slope environments on the eastern side of the Antler foredeep. McGowan Creek–Grandview Canyon section is the most basinal locality in the Antler foredeep, which has most of the Devonian to Upper Mississippian stratigraphy exposed.


A
McGOWAN CREEK-GRANDVIEW CANYON, ID
DEVONIAN
MISSISSIPPIAN
PENN
DEPTH IN METERS
Lower Jefferson Fm.
Grandview Dolomite
Lower McGowan Ck. Fm.
Upper McGowan Creek Fm.
Middle Canyon Fm.
Scott Peak Fm.
South Creek Fm.
Surrett Canyon Fm.
Bluebird Mountain Fm.
B
TENDOY MTS.
ts+minwd
ts+maxwd
Jefferson Fm.
Three Forks Fm.
Paine LS
Middle Canyon Fm.
McKenzie Canyon Fm.
Quadrant SS
Conover Ranch Fm.
Lombard LS
Kibbey SS
C
ASHBOUGH CANYON, MT
ts-minwd
ts-maxwd
Jefferson Formation
Three Forks Fm
Lodgepole Formation
Mission Canyon Formation
Snowcrest Range Group
D
LOGAN, MT
Jefferson Fm.
Maywood Fm.
Three Forks Fm.
Lodgepole Fm.
Mission Canyon Fm.
Snowcrest Range Group
E
SACAJAWEA PEAK, MT
ts+wdmin
ts+wdmax
Jefferson Fm.
Maywood Fm.
Three Forks Fm.
Lodgepole Fm.
Mission Canyon Fm.
Alaska Bench Fm.
Tyler Fm.
Otter Fm.
Kibbey SS
F
MONARCH, MT
ts+wdmin
ts+wdmax
Jefferson Fm.
Maywood Fm.
Three Forks Fm.
Lodgepole Fm.
Mission Canyon Fm.
Alaska Bench/
Tyler Fm.
Heath Fm.
Otter Fm.
Kibbey SS
-380 -370 -360 -350 -340 -330 -320 -310 Ma

events in the Antler orogen. From earliest Famennian to middle Kinderhookian time subsidence across the entire foreland slowed down from previous Frasnian rates, suggesting a relatively quiescent period with little thrust movement in the Antler accretionary wedge. However, the subsidence histories of the proximal foredeep and distal foreland are complicated and difficult to interpret on a regional scale. The numerous unconformities that separate the various members of the Famennian Three Forks Formation have highly variable chronostratigraphic relationships, even over short distances (tens of kilometers) across the study area (fig. 3B). These complex Famennian unconformity relationships cannot be explained by forebulge migration, assuming either a simple elastic or viscoelastic model [cf. Tankard (1986)]. Instead, Famennian unconformity relationships suggest the combined effects of short-term sea-level changes ($\sim 10^6$-yr time scales; Johnson et al., 1985), flexural effects (including forebulge migration), and differential subsidence. The differential subsidence across the foreland may reflect a complex response to (1) episodic loading and intervening relaxation (?) of the foreland lithosphere, (2) reorientation of the principal compressive stress direction from Late Devonian to earliest Mississippian time (suggested by changes in the location of Antler foredeep depocenters; fig. 4) and concomitant changes in deformation patterns in the foreland, (3) variations in the scale or geographic location of thrust loads, (4) independent tectonic response of fault-bounded blocks in the foreland lithosphere, and (5) tectonically induced crustal weakening and development of a low-viscosity zone in the lower crust of the foreland [cf. Howell and van der Pluijm (1990)]. These various models are discussed in greater detail later.

Another major pulse of subsidence affected the Antler foredeep and distal foreland platform in middle Kinderhookian time (figs. 7A–F). Within the limits of time resolution provided by the biostratigraphy, this major subsidence event appears to have been *simultaneous* across the entire study area. Again, however, the instantaneous response across this wide foreland area may not be entirely a flexural response to thrust loading. Flexural wavelengths of the order of 800–1,000 km (500–620 mi) (palinspastic distance from proximal foredeep in Idaho to distal foreland platform in central Montana) are not likely without assuming unrealistically large flexural rigidities for foreland lithosphere. There is additional stratigraphic and structural evidence that the foreland further to the east of our study area was affected at the same time (McCabe, 1954; Clement, 1986; Gerhard et al., 1987; LeFever et al., 1987). Dip-slip and reverse movement on high-angle faults in central and easternmost Montana during the Late Devonian to Late Mississippian has been documented (Plawman, 1983; Clement, 1986; Nelson, personal communication, 1990). The location of some of these distal cratonic structures requires a tectonic response over 1,000 km (620 mi) (palinspastic) from the proximal foredeep.

The stratigraphic record of this pulse of subsidence is the onlapping Lodgepole Formation and its more basinal stratigraphic equivalents. Coincidence of this pulse of subsidence with an inferred eustatic sea-level rise (Sandberg et al., 1983) produced the Cottonwood Canyon Member, the thin condensed sequence at the base of the Lodgepole Formation. The coincidence of this subsidence event and the inferred eustatic rise partly but not completely explains the apparent widespread tectonic response of the foreland lithosphere (discussed later). By the beginning of Osagean time subsidence had slowed down across the foreland. This decrease in subsidence rate allowed progradation of the Mission Canyon platform (Reid and Dorobek, 1989, 1991; Reid, 1991).

Another pulse of subsidence, although less pronounced than the middle Kinderhookian pulse, affected the entire study area in middle to late Meramecian time. However, within the limits of biostratigraphic resolution the onset of this pulse of subsidence apparently was not simultaneous across the study area. One reason for the apparently diachronous onset of Meramecian subsidence may be the regional unconformity on top of the Mission Canyon platform. This unconformity developed during a eustatic (?) sea-level lowstand and resulted in extensive erosion and karstification of the Mission Canyon platform (Sandberg et al., 1983; Ross and Ross, 1987; Sando, 1988; Reid, 1991). Regional biostratigraphic and lithostratigraphic relationships also suggest that some deformation of the foreland platform occurred during development of the post–Mission Canyon unconformity (Sando et al., 1975; Sando, 1988). The foreland platform may have had as much as 60–100 m (200–330 ft) of local relief produced by the combined effects of deformation, erosion, and karstification before marine flooding in the middle to late Meramecian [cf. Sando (1988)]. Therefore the apparent onset of subsidence across the platform was dependent on the time of initial marine flooding across an unconformity surface with significant paleotopography; paleohighs would have been flooded at a later time than paleovalleys, and subsidence apparently would have begun later on the paleohighs.

Meramecian subsidence continued on to the Early Pennsylvanian. This episode of subsidence was gradual and appears to have been nearly constant across the foreland platform. This gradual platform subsidence allowed deposition of shallow marine to marginal marine facies. Foredeep subsidence, however, occurred at a slightly higher rate than typical platform subsidence rates, as indicated by the greater slopes on the Meramecian–Pennsylvanian segments of the McGowan Creek–Grandview Canyon subsidence curve. Foredeep deposits during this time included thick open marine carbonates of the White Knob Group in eastern parts of the basin (Huh, 1967, 1968; Mamet et al., 1971) and westerly derived, proximal siliciclastic facies of the upper Copper Basin Formation in western parts of the foredeep (Paull et al., 1972; Nilsen, 1977; Skipp et al., 1979). Another episode of Antler thrust movement probably occurred in Late Mississippian time, providing the load for foredeep subsidence and the source for coarse-grained siliciclastic facies of the upper Copper Basin Formation.

**Eustatic versus tectonic signals in the subsidence curves** It is important to note that there is a eustatic component in the subsidence curves from Montana and Idaho. As discussed earlier, the Frasnian–Famennian pulse of platform subsidence may reflect, at least partly, a Late Devonian eustatic sea-level rise (Bond and Kominz, 1991), whereas time-equivalent foredeep subsidence reflects both eustatic rise and tectonically driven subsidence. The apparent simultaneous nature of Mississippian inflection points on the subsidence curves also may be due partly to eustatic sea-level fluctuations. Previous studies have suggested time-equivalent eustatic rises that *approximately* coincide with pulses of increased subsidence on the curves from Montana and Idaho (Hallam, 1984; Johnson et al., 1985; Ross and Ross, 1987; Bond et al., 1989). However, Bond and Kominz (1991) have estimated that the total amount of sea-level rise from Late Devonian (base of Frasnian) to Middle Mississippian (lower Chesterian) time was of the order of 180 m (590 ft) [±35 m (114 ft), maximum error over entire time span], based on subsidence analyses of midcontinent strata in Iowa. The Iowa section was used to estimate eustatic change because it was deposited near the center of the North American craton, an area that presumably was tectonically stable during the Late Devonian to Middle Mississippian. The 180-m (590-ft) eustatic rise during this interval is distributed as follows: (1) Frasnian to top of Famennian, ~100 m (~330 ft); (2) base of Kinderhookian to top of Osagean, ~55 m (~180 ft); and (3) base of Meramecian to lower Chesterian, ~25 m (~80 ft). The short-term incremental pulses of subsidence in Mississippian parts of the curves from Montana and Idaho (which include eustatic fluctuations) greatly exceed these estimates of incremental sea-level rise. This holds true across the entire study area and suggests that the Mississippian pulses of subsidence probably largely reflect true tectonically driven subsidence. In addition, the magnitude of subsidence during any interval of subsidence varies across the study area. Subsidence also is always greater in the proximal foredeep than on the adjacent platform, although subsidence does not appear to decrease monotonically away from the foredeep, as a simple flexural model would predict. These regional variations in subsidence cannot reflect changes in eustatic sea level but instead must be related to differential tectonic subsidence.

## Controls on differential subsidence across the Montana–Idaho foreland

In this section we attempt to relate the different types of loading associated with the Antler orogeny, the structure of the pre-Antler foreland lithosphere, and the timing of Antler events to develop conceptual kinematic models for differential subsidence across the study area.

**Possible basement controls on differential subsidence** Highly variable lithologies (metasedimentary to layered mafic intrusives), metamorphic grade (greenschist to amphibolite), terrane ages (Archean to late Proterozoic), and structural domains are exhibited in the scattered exposures of Precambrian basement rocks across the study area. The lithologic variation in Precambrian crustal rocks may reflect heterogeneity and variable rheologies in underlying lithosphere, as inferred by, for example, McMechan (1981), Winston et al. (1984), and Hoffman (1989). This lithospheric heterogeneity may have produced some of the differential response in the Antler foreland because variable lithosphere rheologies might respond differently to tectonic loading.

Variations in basement lithology are likely to be of secondary importance compared to the foreland response caused by imposing a regional stress field on preexisting basement structures. The geographic coincidence of Antler foreland structures with Precambrian structures suggests reactivation of Precambrian faults during Antler convergence.

**Flexure versus in-plane stress** Continental lithosphere beneath the Devonian–Mississippian cratonic platform across Montana most likely was thick and cold but had been segmented during at least two major pre-Antler extensional events in the middle Proterozoic and late Proterozoic–Early Cambrian (Hoffman, 1989; Oldow et al., 1989). Incipient Antler convergence in the Early to Middle Devonian apparently began far outboard of the lower Paleozoic passive margin hinge zone, and flexural bending may not have been transmitted very far inboard because of the distal position of the load and/or the possible low flexural rigidity of the loaded lithosphere. Frasnian sedimentation in Montana apparently began only after the Antler thrust load moved far enough eastward so that flexure involved lithosphere beneath westernmost parts of the Jefferson platform or possibly because of a eustatic sea-level rise that flooded large parts of the North American continental interior (Bond and Kominz, 1991).

Flexural bending also could explain some of the Late Devonian–Early Pennsylvanian differential subsidence across the foreland platform if the Antler accretionary wedge was areally limited throughout the course of the Antler orogeny and individual, mechanically independent segments of foreland lithosphere were loaded (fig. 8A). Antler collision along the continental margin in Idaho probably was diachronous during Late Devonian to Early Pennsylvanian time, as suggested by the southeastward migration of Antler foredeep depocenters (fig. 4). Diachronous collision also is suggested by the apparent southeastward migration of western source areas for synorogenic siliciclastic sediment depos-

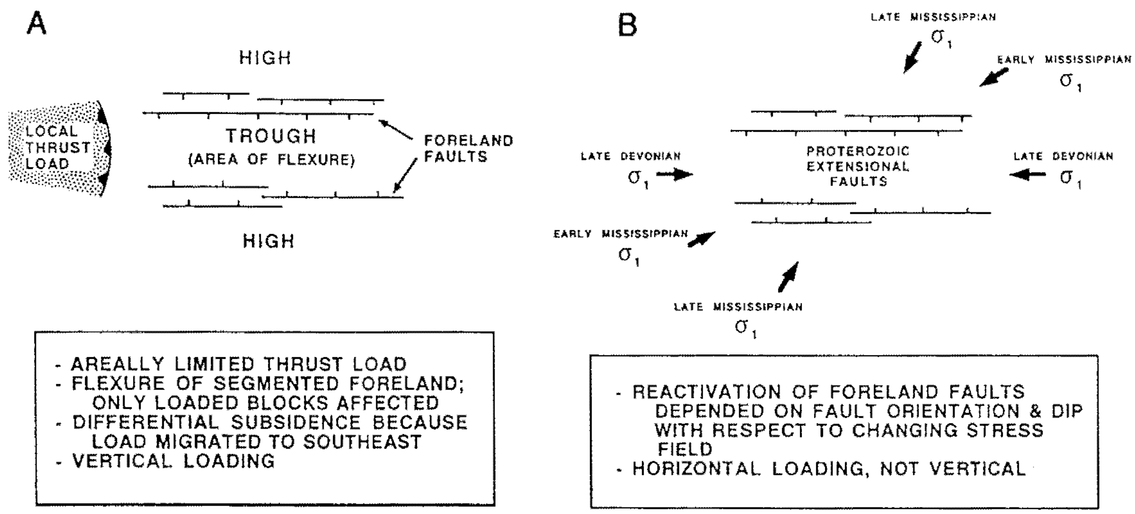



**Figure 8.** Generalized conceptual kinematic models explaining the differential subsidence across the Antler foreland in Montana and Idaho. (A) Differential subsidence resulting from subregional vertical loading. Emplacement of areally limited thrust loads on broken foreland lithosphere may have caused flexural response of mechanically independent foreland blocks. See text for more complete discussion of model. (B) Differential subsidence resulting from regional horizontal loading. Transmission of in-plane compressive stress associated with Antler convergence may have caused movement on old Proterozoic faults. Amount of movement and sense of shear along ancient fault zones would have depended on orientation of faults with respect to regional stress field. Migration of Antler foredeep depocenters from Late Devonian to Late Mississippian time suggests that orientation of principal stress direction (indicated by arrows) also may have changed through time. However, actual stress field orientation at any time would have depended on the geometries along the margins and the relative motions of the North American plate and colliding Antler terrane. Such data are not available at this time. Nonetheless, individual faults and fault zones apparently had different responses through time as Antler convergence progressed.

ited in the Antler foredeep from Devonian to Pennsylvanian time (Nilsen, 1977; Skipp et al., 1979). Therefore thrust loads probably were not distributed evenly along the ancient continental margin through time, producing *subregional vertical loads* that changed position as Antler collision progressed along the margin. Individual blocks of the segmented Antler foreland, especially those bounded by east-west-trending Proterozoic faults, may have responded separately and sequentially to the migrating thrust load maxima. Similar foreland behavior is exhibited in the Apennine foreland system of Italy, where the foreland basin area is distinctly compartmentalized into subbasins because the foreland lithosphere apparently is segmented by tear faults that trend at high angles to the thrust load (Royden et al., 1987). Each foreland segment is mechanically independent of adjacent segments, producing offset foreland depocenters with separate forebulge areas.

However, the differential subsidence of the Montana platform cannot be attributed entirely to lithospheric flexure in response to vertical loading, even if the lithosphere beneath the platform was segmented by high-angle faults. Much of the Montana platform was located 500–1,000 km (310–620 mi) (palinspastic distance) inboard from the deepest part of the Antler foredeep. These distances are greater than flexural wavelengths produced by assuming geologically reasonable rheologies for loaded plates. Therefore the potential for flexural response of the segmented foreland lithosphere would have been constrained to Idaho and westernmost Montana [based on maximum flexural wavelengths of 400 km (250 mi) and the palinspastic reconstructions of Peterson (1986) and Sandberg et al. (1983)]. In addition, many of the paleostructures across the cratonic platform were oriented at high angles to the axis of the Antler foredeep and the inferred strike of the Antler thrust belt; the strike of flexural features on the cratonic platform should be subparallel to the Antler foredeep–thrust belt. Finally, many of the foreland structures had relatively short wavelengths [often less than 100 km (60 mi)]. Multiple short-wavelength foreland structures that are oriented at high angles to the axis of a foredeep are not predicted for distal foreland areas by simple flexural models (i.e., horizontal loads not considered).

An alternative explanation for the differential subsidence across the foreland is the reactivation of Proterozoic faults by in-plane compressive stress during Antler convergence. Stress can be transmitted laterally over hundreds of kilometers inboard from continental plate margins (Gay, 1980; Zoback and Zoback, 1980; Zoback et al., 1985; Cloetingh and Wortel, 1986; Craddock and van der Pluijm, 1989). Inversion tectonics involving reversals of movement along high-angle faults are well documented in the Alpine foreland, over 1,000 km (620 mi) from the leading edge of Alpine thrusts (Ziegler, 1987a,b). These movements might be due to in-plane stresses that reactivated preexisting fault systems (Karner, 1986; Cloetingh, 1988). Similar reversals in dip-slip movement on high-angle faults that cut through Devonian and Mississippian strata in eastern Montana have been documented from borehole data and on seismic lines (Plawman, 1983; Clement, 1986; Nelson, personal communication, 1990). It is difficult to explain this deformation without calling on a regional stress field produced by tectonic loading along boundaries of the North American plate.

The actual response of Proterozoic faults across the Antler foreland to the hypothesized regional stress field would have depended on fault orientation, mechanical properties along fault surfaces, and the orientation of the regional stress field. As the Antler foredeep depocenters and inferred locations of thrust load maxima migrated to the southeast from Late Devonian to Early Pennsylvanian time, the orientation of the principal stress direction also may have changed with respect to the east-west-trending foreland faults. As the principal stress direction became more oblique relative to the preexisting foreland faults, strike-slip and/or dip-slip movement may have occurred along the faults (fig. 8B). However, the actual orientations of any possible changes in the Devonian–Mississippian regional stress fields cannot be resolved more accurately without better constraints on the actual geometries of the colliding plate boundaries and possible changes in relative plate motions *during* Antler convergence.

It also is possible that the *magnitude* of in-plane stress in the distal foreland lithosphere varied during the Antler orogeny. If this is true, then episodic changes in the magnitude of horizontal in-plane stress acting on the preexisting faults in the distal foreland also may have caused some component of differential subsidence in that area [cf. Cloetingh (1988)].

Elevated levels of in-plane compressive stress also might have produced a zone of lower effective viscosity in the lower crust, as suggested by Howell and van der Pluijm (1990). During periods of increased horizontal loading, the elevated in-plane stress (possibly coupled with anomalously high pore fluid pressures) may have lowered the effective viscosity in lower crustal levels, making the lower crust unable to support upper crustal loads (Howell and van der Pluijm, 1990). In addition to the decreased viscosity of the lower crust, a large preexisting excess mass in the upper crust is necessary to produce subsidence during episodes of peak horizontal loading (Howell and van der Pluijm, 1990). Regional gravity anomalies in the study area are associated with Mesozoic–Cenozoic plutons (Hanna et al., 1989). The present distribution of gravity anomalies neither proves nor disproves the existence of an excess mass in the upper crust of Montana and Idaho in Devonian–Mississippian time. Regardless, Howell and van der Pluijm's model does not adequately explain the differential subsidence and, more important, the tectonic inversion across the study area during Antler convergence.

In summary, reactivation of preexisting foreland faults by *regional horizontal loading* might account for some of the short-wavelength, variably oriented structures across the Antler foreland. Relative motion along fault surfaces most likely would have depended on fault orientation with respect to a changing regional stress field, mechanical characteristics of the faults, or the episodic nature of horizontal loading of the faults.

## Implications for other foreland settings

**Timing of Antler convergence events through analysis of foreland stratigraphy** Our approach has been to examine the response of the distal foreland area to episodic convergence through detailed study of the foreland stratigraphy. Preexisting basement structures apparently had significant influence on the response of the distal foreland. This study suggests that approximate dates for episodic convergence events can be derived from detailed stratigraphic studies and subsidence analyses of foreland strata, even in areas where proximal foredeep facies and ancient thrust belt complexes are poorly preserved [cf. Jordan et al. (1988)]. In fact, episodic convergence events might be identified through analyses of distal foreland strata that were deposited hundreds of kilometers inboard from foredeep depocenters, so long as most of the accommodation space in the distal foreland is produced by tectonic subsidence and not just by eustatic sea-level rise. Subtracting the estimated increments of eustatic sea-level rise during the Frasnian to Meramecian (Bond and Kominz, 1991) from the Idaho–Montana subsidence curves indicates that most of the pulses of apparent subsidence in the Antler foreland have a large component of remaining tectonic subsidence. This is true at least for Mississippian parts of our subsidence curves but is more ambiguous for Frasnian–Famennian segments of platform subsidence curves, when some of the apparent subsidence is similar in magnitude to estimates of Frasnian–Famennian eustatic sea-level rise. The episodic foreland subsidence therefore probably is a true indicator of episodic convergence and/or episodic changes in the in-plane stress field during much of Antler time.

**Implications for forward models of carbonate platform evolution** Carbonate platform studies that incorporate quantitative subsidence analyses typically are from passive margin or isolated platform settings [e.g., Winterer and Bosellini (1981), Freeman–Lynde and Ryan (1987), and

Bond et al. (1989)]. Two-dimensional forward models of carbonate platform evolution also incorporate subsidence into the model results [e.g., Koerschner and Read (1989), Scaturo et al. (1989), and Read et al. (1990)]. However, carbonate sequences from passive margins or isolated platforms often are selected for forward modeling because subsidence in these tectonic settings is assumed to follow reasonably predictable patterns.

In contrast, factors that affect the evolution of carbonate sequences from foredeeps or adjacent foreland settings are more difficult to constrain. A large source of uncertainty is the subsidence history of the foredeep basin and adjacent foreland. The flexural response of the thrust-loaded lithosphere may not always be easily predictable because of uncertainties in the timing and scale of thrust-loading, possible subcrustal loads and changing boundary conditions, and unknown rheologic properties of the thrust-loaded lithosphere (Beaumont, 1981; Jordan, 1981; Royden and Karner, 1984; Schedl and Wiltschko, 1984; Stockmal et al., 1986; Stockmal and Beaumont, 1987; Beaumont et al., 1987; Jordan et al., 1988). Response of the foreland to thrust loading may be even more complicated if the foreland lithosphere is lithologically heterogeneous or broken by preexisting faults that penetrate deep into the lithosphere (Karner, 1986; Ziegler, 1987a,b; Royden et al., 1987).

This study illustrates a well-documented example of several carbonate platform sequences that were deposited in distal foreland areas and were affected by episodic convergence events. The subsidence histories in similar foreland settings might be highly variable in time and space. Therefore caution should be exercised before applying one- or two-dimensional models of carbonate sedimentation in similar tectonic settings.

## Summary

In this study we have documented the response of a foredeep and adjacent cratonic platform to episodic convergence events along a continental margin located hundreds of kilometers outboard of the foreland area. The complex chronostratigraphic and lithofacies relationships across the Antler foreland of Montana and Idaho provide a good example of the need for models that incorporate both flexure and the effects of horizontal in-plane stress to explain the actual foreland stratigraphy. Although we did not test an actual flexural model or an in-plane stress model in this area, the Montana–Idaho foreland may be an ideal place to examine the relative importance of horizontal versus vertical tectonic loads on foreland lithosphere. However, the following conclusions generated by this study may have application to other studies of foreland behavior.

1. Mechanical discontinuities, such as deep-seated preexisting faults, may influence the tectonic response of foreland areas during convergence events. More important, preexisting faults in foreland lithosphere may cause essentially unpredictable differential subsidence, which can have a dramatic effect on foreland stratigraphy.

2. The patterns of regional subsidence described here may occur in other foreland areas that (a) occur above old, thick continental lithosphere that is segmented by deep preexisting faults and (b) are far inboard of plate margins where the actual convergence occurred.

3. Flexural models that incorporate only vertical loading cannot account for all the differential subsidence that occurred across the Montana–Idaho foreland area during Antler convergence. The transmission of in-plane stress associated with convergence and the effect of in-plane stress on preexisting foreland faults may have an important additional effect on differential subsidence, especially in distal foreland areas.

4. It may be possible to accurately constrain the timing of emplacement of major thrust loads by examining the subsidence history of *distal* foreland areas if the magnitude of eustatic sea-level variation can be estimated independently. This approach may be useful in structurally complex terranes where the thrust belt, which records the deformation history of the convergence events, has been destroyed by erosion or obscured by later deformation.

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