General Stratigraphy and Regional Paleotectonics of the Western Montana Overthrust Belt[1]

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A composite stratigraphic section ranging in age from late Precambrian to Holocene is present in the western Montana overthrust belt. Paleozoic rocks are 2000 to 5000 m thick and are dominated by shallow marine shelf limestone and dolomite facies, much of which is of carbonate bank or reefal origin in the lower and middle Paleozoic, and by shelf marine sandstone and carbonate in the upper Paleozoic. Source areas for Paleozoic and early Mesozoic clastic facies were in south-central Canada and north-central United States. Mesozoic rocks between 300 m and 7000 m thick are dominated by shallow water marine clastic facies in the pre-Cretaceous and by continental and near shore marine facies in the Cretaceous section, which becomes progressively coarser, volcanic-rich, and more continental in the younger beds. The source area for Mesozoic clastics was in east-central and northern Idaho and westernmost Montana which, beginning in Late Jurassic time, became the site of increasingly intense tectonic growth that culminated in the development of the thrust and fold belt and associated igneous activity in Middle to Late Cretaceous and early Tertiary time.

The main structural framework of western Montana was developed during late Precambrian Belt Supergroup deposition, and many of the prominent paleostructural elements persisted through most of the remainder of geologic time. The more important of these that influenced the nature and distribution of sedimentary facies during Paleozoic and Mesozoic time are the Lemhi arch, Alberta shelf, Beartooth shelf, Belt Island complex, Boulder high, Coeur D'Alene-central Montana trough, Big Snowy trough, and Snowcrest trough. In east-central Idaho, the Muldoon trough (north segment of the Sublett basin), which in Paleozoic time lay between the Antler orogenic belt and the Montana shelf province, was a regional area of active subsidence that received a great thickness of Paleozoic shallow to deep water marine sediments.

The lower and middle Paleozoic rock facies in western Montana includes a large thickness of porous dolomite, and the upper Paleozoic contains a large volume of clean shelf sandstone, some of which has good porosity. Potential petroleum source rocks are present in the Devonian, Mississippian, and Permian beds, some of which, despite deep burial of most of the Paleozoic section, are not highly altered thermally. The Mesozoic rock facies is primarily continental in origin, but a broad belt of intertonguing near shore marine and continental facies is present. The sandstone bodies in this facies offer reasonable potential for significant biogenic and thermal gas accumulations under adequate trapping conditions. Tertiary continental and lacustrine beds, as much as 3000 m or more thick, were deposited in localized downwarped basins. These beds contain substantial thicknesses of discontinuous alluvial sandstone and some carbonaceous to coaly beds and are of interest for their biogenic gas potential.

INTRODUCTION

This paper reviews the regional tectonic and stratigraphic history and the regional depositional facies relationships of Precambrian through Holocene rocks in western Montana (Figure 1). It is not intended to be a comprehensive or detailed description of the complete stratigraphic section. The reader is referred to other publications listed in the references for details of specific stratigraphic units, faunal data, age indications, regional correlations, and reviews of previous work.

The maps and charts included in this paper were prepared on a palinspastic base designed to restore stratigraphic data to approximate depositional positions. These palinspastic reconstructions use modifications of major thrust displacements estimated by several authors, including Mudge (1970), Skipp and Hait (1977), Ruppel (1978), Ruppel and Lopez (1984), and W. J. Perry (personal communication, 1980). Positions of the major thrust zones and the displacements used are shown in Figures 2 and 3. Figures 4, 5, 6, and 7 are palinspastically restored lithofacies cross sections.

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[1] Reprinted with minor modification and updating of references from Symposium Guidebook to Southwest Montana Geological Society, 1981, Billings, Montana, p. 5-35, with permission of the Montana Geological Society.

Stratigraphic data were compiled from a variety of sources, including personal files; numerous unpublished theses; publications of the U.S. Geological Survey, the Montana Bureau of Mines and Geology, and the Montana Geological Society; and borehole data from geophysical and lithologic logs of the American Stratigraphic Company. Most of the important data sources are listed in the references. Paleozoic stratigraphic intervals for map construction were selected primarily on the basis of the predominance of carbonate versus clastic facies. Mesozoic intervals, which are primarily clastic deposits, were selected on the basis of major differences in the nature and depositional environments of the clastic sediments.

REGIONAL STRUCTURAL SETTING

The surface geology of western Montana is dominated by the western overthrust and disturbed belt, a northwestern area of primarily Belt Supergroup exposures, a west-central region of Mesozoic intrusives, a large east-central area of late Mesozoic and early Tertiary volcanic and coarse conglomeratic deposits, and a south-central area of exposed older Precambrian rocks. McMannis (1965) separated the region into three general provinces on the basis of differences in the tectonic and stratigraphic character of the rocks. First, the *Belt province* is characterized by exposures of the thick Precambrian Belt Supergroup. No exposures of older Precambrian rocks are known in this province, and several northwest-southeast-trending vertically faulted troughs filled with middle and late Tertiary sediments are present within this province. Second, the *batholithic province* is mainly confined between the Montana lineament and the west-east "Perry line" of faults north and west of Bozeman. The Boulder and Idaho batholith complexes and related intrusives and the "Sapphire block" (Hyndman et al., 1975) are within this province, as are several sharply defined Tertiary basins, many containing thick nonmarine deposits. Exposures of older Precambrian rocks are absent or rare within this province. Third, the *basement province* corresponds closely to the position of the Beartooth shelf (Figures 1, 2, and 3) and is dominated by exposures of older Precambrian metamorphic rocks, with Belt Supergroup rocks generally being absent. Paleozoic and Mesozoic sequences are thinner here than in the other provinces. It contains several well-defined block-faulted Tertiary basins and a network of high-angle faults.

The Montana plains province of eastern and central Montana, which is characterized by less complex structure and more uniform stratigraphy, merges westward into the northern part of the disturbed belt, the eastern part of the basement province, and the eastern flank of the Crazy Mountain basin.

These subdivisions of this complex area are reasonably well defined, although some characteristics of any given province are also found within the others.

REGIONAL PALEOGEOGRAPHY AND PALEOSTRUCTURE

Paleozoic rocks (Figure 2) range in thickness from 3000 ft (900 m) or less in the Montana plains province to over 15,000 ft (4500 m) in southwestern Montana (Figure 2). In the central parts of the Muldoon trough or Sublett basin of Idaho, they are at least 50,000 ft (15,000 m) thick (Peterson, 1977). The influence of several persistent Paleozoic plaeostructural features is reflected in thickness patterns as well as facies patterns of the Paleozoic sequence.

Mesozoic rocks (Figure 3) are absent over much of western Montana, which during most of this time was an actively emergent source area for clastic debris. Thickness patterns of Mesozoic rocks, when restored, indicate that they were probably more than 20,000 ft (6000 m) thick in parts of westernmost Montana but thinned eastward to less than 5000 ft (1500 m) in parts of the Montana plains (Figures 3 and 6). Several of the main Paleozoic paleostructural elements persisted throughout Mesozoic time, although development of the overthrust belt and intrusive activity changed the nature of paleostructural activity from one of relatively gentle uplifts and downwarps to one of greatly increased intensity during Mesozoic time.

During Paleozoic and early Mesozoic time, western Montana was a part of the broad craton–miogeosyncline border zone that extended the length of the western North American continent (Figure 1). Transgressive marine carbonate deposits dominated sedimentary facies during most of this time, but there were frequent regressive interruptions when clastic marine sediments, primarily originating from source terranes to the east and northeast, spread over previously deposited shallow water carbonate beds. Major clastic source areas in the western United States during Paleozoic time were located in and near the Canadian shield and the Transcontinental arch to the east. After Middle Devonian time, the Antler orogenic belt to the west became an active source area, although rarely were clastic materials from this belt of uplifts transported eastward far enough to invade the Rocky Mountain shelf.

Stratigraphic facies and thickness studies document the presence of several positive and negative paleostructural areas with varying degress of persistence that affected the depositional patterns of the major stratigraphic units. Some of these were present during late Precambrian Belt deposition and many continued throughout most or all of Phanerozoic time (Figures 1, 2, and 3). Among the more persistent regional features were the following (refer to the thickness and lithofacies maps of late Precambrian through Cretaceous (Montanan) sedimentary rocks in Figures 8 through 18):

The *Wyoming shelf* (Figure 1) occupied most of the Wyoming and southernmost Montana area during Paleozoic and early Mesozoic time.

The *central Montana trough* (Figures 1, 2, 3, and 7) was a subsiding trough area that was active during Precambrian Belt time and persisted with varying degrees of intensity throughout the remainder of geologic history. This trough area was dominated by carbonate, evaporite, and fine-grained clastic deposition during Paleozoic time and by thick clastic sediments, partly intertongued with volcanic sediments in Mesozoic time.

The *Alberta shelf* (Figures 1, 2, and 3) bordered the central Montana trough on the north and occupied the site of present-day southern Alberta, western Saskatchewan, and north-central Montana. This province was the site of predominantly carbonate and fine-grained clastic deposition

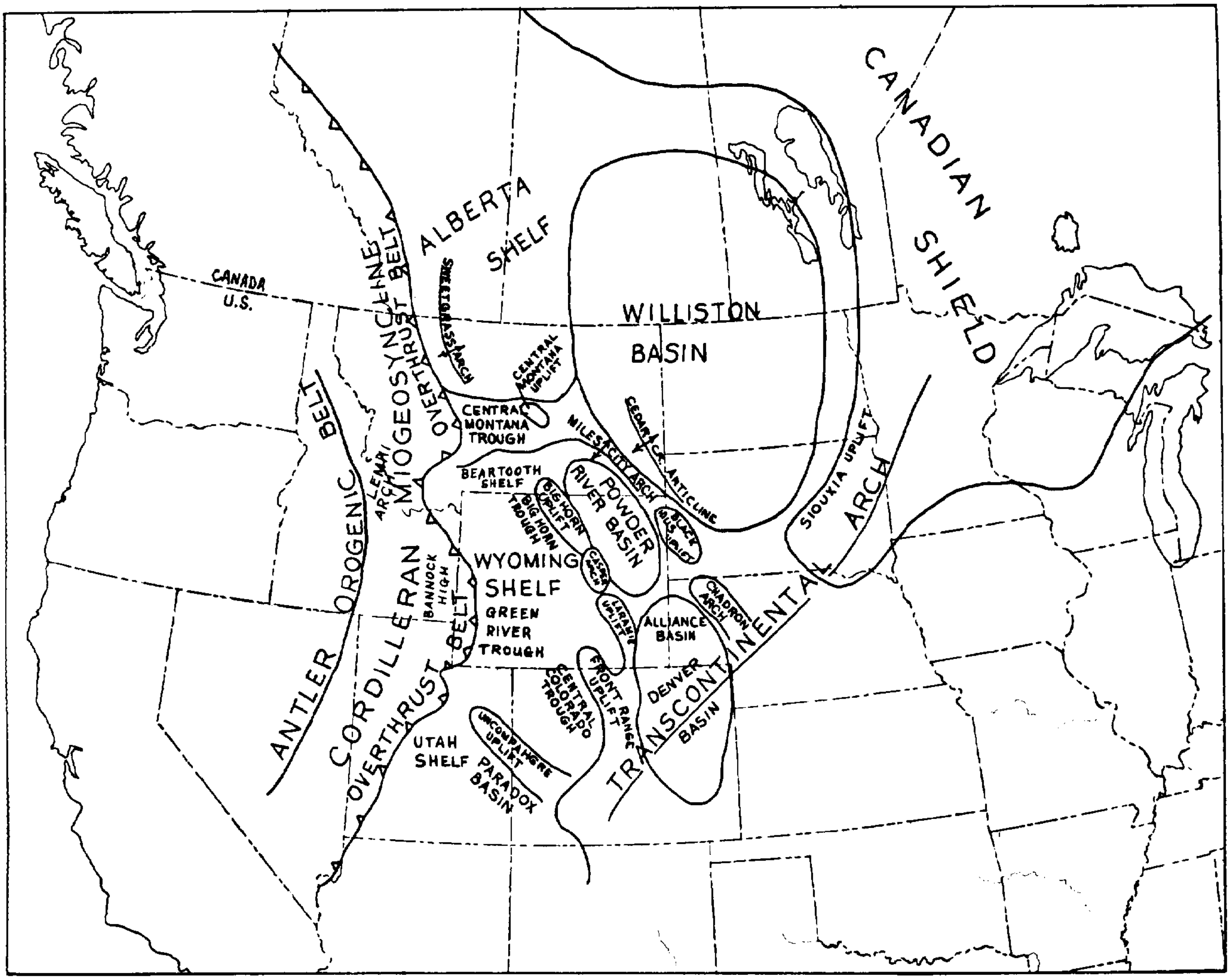

Figure 1—Regional paleogeography and paleostructure of the Rocky Mountain region during Paleozoic and Mesozoic time. Modified after Peterson (1981a).

in Paleozoic time and shelf sand, clay, and minor carbonate sediment in Mesozoic time. The Sweetgrass arch is a paleostructural element of the Alberta shelf.

The *Coeur D' Alene trough* (Figures 2, 8, and 9) was described by Harrison et al. (1974) as an elongate depositional trough during late Precambrian Belt deposition. This feature is part of a west-central downwarp that includes the central Montana trough. Restored thickness maps record evidence of its persistence through at least Paleozoic time and perhaps part of Mesozoic time.

The *Muldoon trough* (Figure 2) was a strongly subsiding Paleozoic trough area that formed the northern segment of the Sublett basin of southeastern Idaho and was particularly prominent in Devonian and Mississippian time (Rose, 1976).

The *Lemhi arch* (Figures 2 and 9) was an active positive area located in southwestern Montana and east-central Idaho. This feature was originally defined by Sloss (1954) as a Devonian arch, but it has since been documented in several other parts of the geologic column. In late Precambrian, Cambrian, and part of Devonian time it was a source area fro clastic sediments and it formed a part of the cratonic shelf margin (Figures 2, 9, and 10) during the remainder of Paleozoic and early Mesozoic time. Several other names have been applied to various parts or phases of paleostructural growth in this area (Beaverhead arch, Scholten, 1967; Salmon River arch, Armstrong, 1975; Tendoy arch, Tysdal, 1976). The Lemhi arch may have extended toward the northwest to include the Precambrian source area referred to as "Belt Island" (Figures 2 and 8) by Harrison et al. (1974).

Other paleostructural features that were more local and less persistent include the following.

The *Beartooth shelf* (Figures 1, 2, and 3) was a northern part of the Wyoming shelf that bordered the central Montana trough on the south and was bounded on its west flank by the Greenhorn fault, which formed the east border of the Snowcrest trough (Figures 2 and 3). The Greenhorn fault underwent several phases of growth during Paleozoic and later time (Hadley, 1980).

The *Big Snowy trough* (Figures 2 and 3) was an elongate west-east belt of active subsidence during Mississippian and Pennsylvanian deposition and probably persisted well into

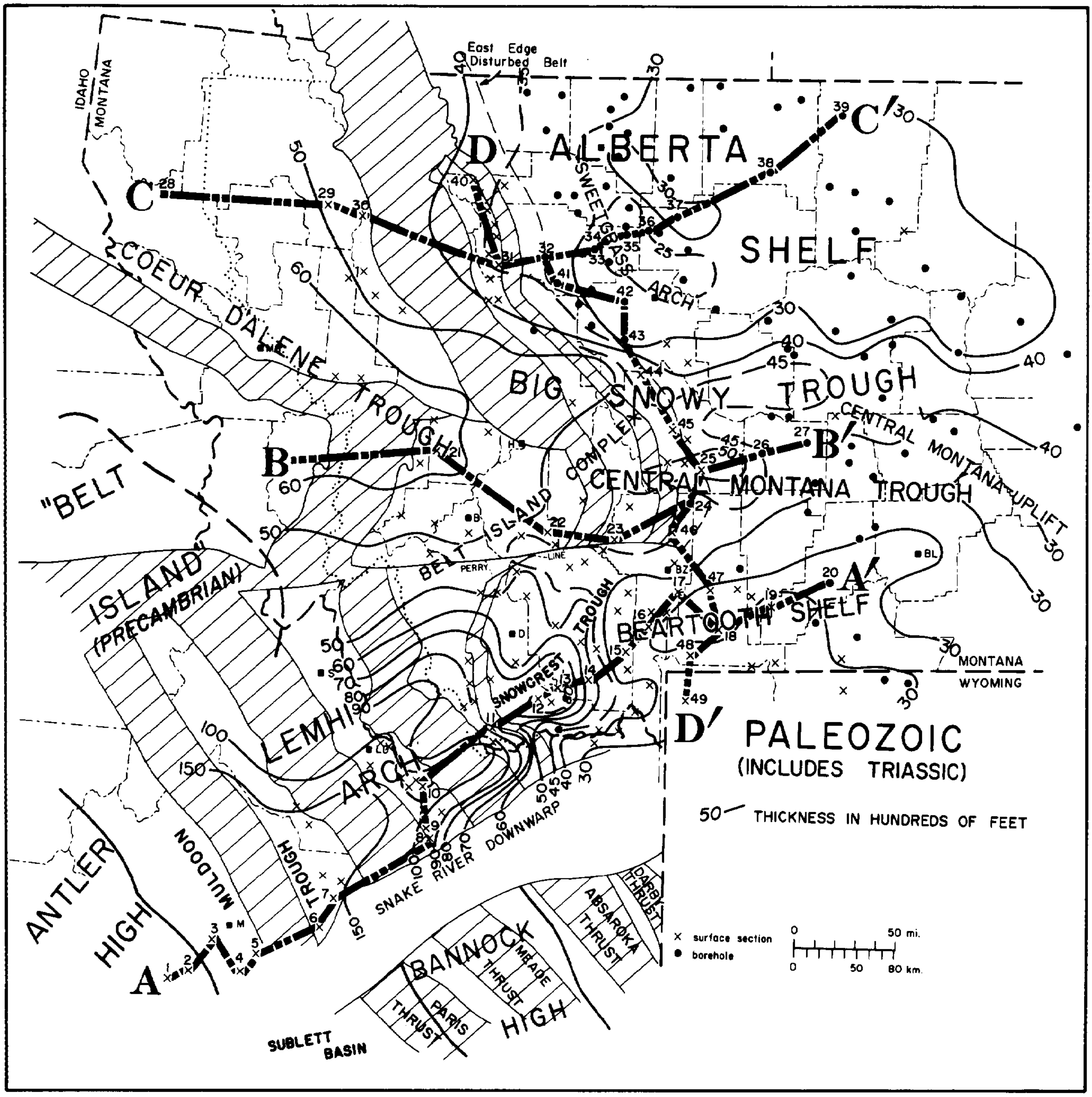

Figure 2—Approximate thickness of Paleozoic rocks, including Triassic, partly restored, showing main paleotectonic features of Paleozoic time. Cross-hatching indicates distances used for palinspastic reconstruction in main thrust area of western Montana and Idaho. Present-day outline of western Montana boundary is shown by dotted line within thrust belt. Lines of cross sections A–A′ through D–D′ in Figures 4, 5, 6, and 7 are shown. Cities shown for Montana are Butte (B), Billings (BL), Bozeman (BZ), Dillon (D), Great Falls (GF), Helena (H), Missoula (MS), and Shelby (SH); and for Idaho are Leadore (L), McKay (M), and Salmon (S). Palinspastic reconstruction of thrust belt features in southeastern Idaho is shown for comparison with those of western Montana. Modified after Peterson (1977, 1980). Used with permission of the Society of Economic Paleontologists and Mineralogists.

Mesozoic time. It formed the more active northern segment of the central Montana trough, which tended to separate into two parallel trough features in late Paleozoic time.

The *central Montana uplift* (Figure 1) was particularly active in Devonian time, but evidence also indicates its persistence during several times in the Paleozoic. Ancestral growth of this feature may have occurred in Cambrian time, as shown by thickness patterns based on relatively limited control (Figure 9).

The *Belt island* system (Figures 3 and 15) was a complex of shallow marine emergent areas identified by Imlay et al. (1948) to explain variations in thickness patterns of marine Jurassic cyclic units. Earlier phases of this complex are evident for late Paleozoic and early Mesozoic time, although

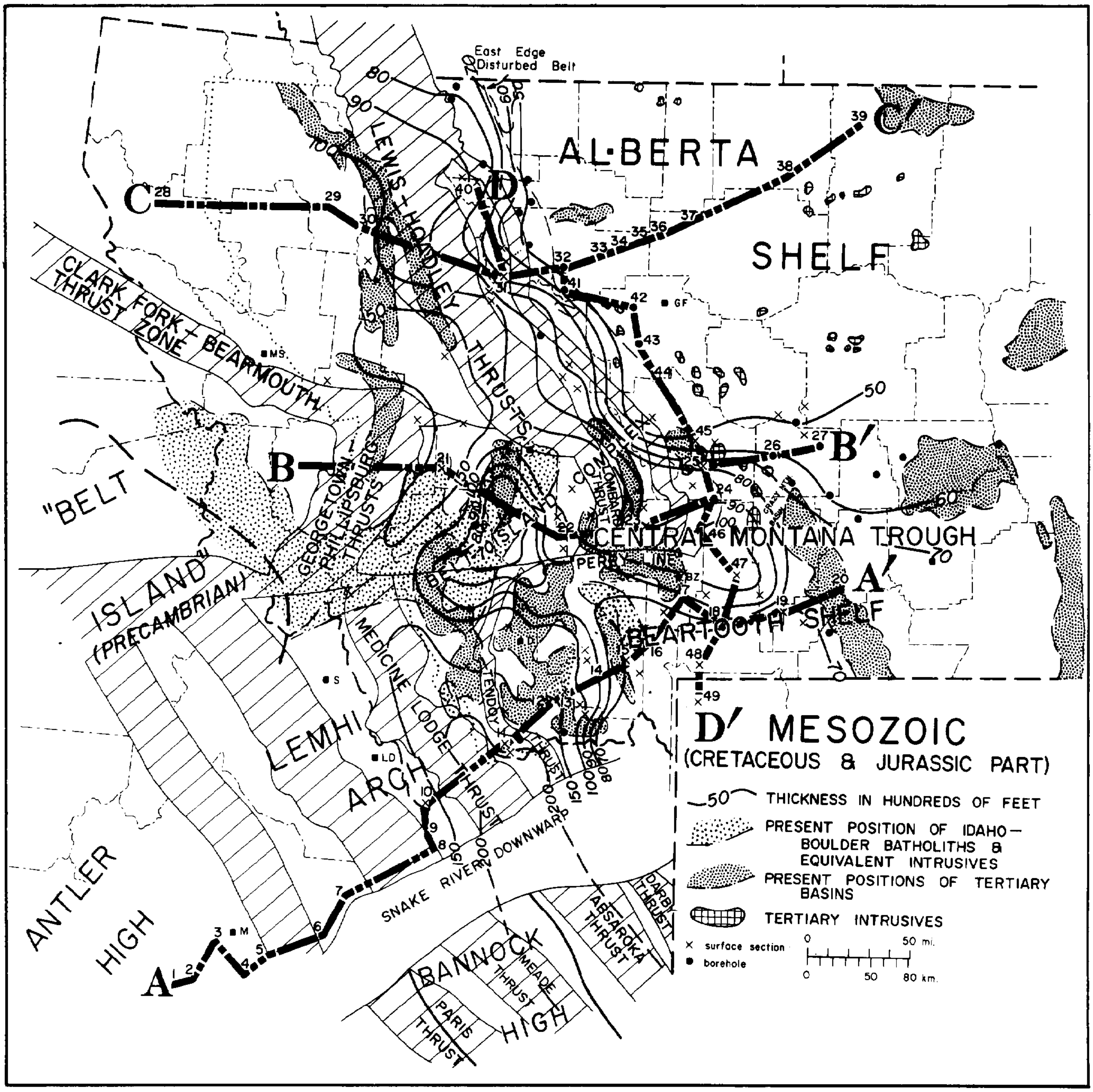

Figure 3—Approximate thickness of Mesozoic rocks, excluding Triassic, partly restored, showing main paleostructural features of Mesozoic time. Cross-hatching indicates distances used for palinspastic reconstruction in main thrust belts of western Montana and Idaho. See Figure 2 legend for further explanation.

removal of a substantial part of the older section by pre-Middle Jurassic erosion prevents an accurate appraisal of its early history. Parts of the Belt island complex may have been involved with the initial phases of structural activity that culminated in some Laramide structures.

The *Boulder high* (Figures 3 and 15) was noted by Mutch (1961) as an area of thinning in Early Cretaceous time. Evidence of its effect is also found in post-Cambrian thickness patterns. This feature formed a part of the Jurassic Belt island complex, and ultimately it became the site of emplacement of the Boulder batholith in late Mesozoic time. The northeasterly orientation of the northwest border of the Boulder high coincides closely with the trend of the Great Falls tectonic zone of O'Neill and Lopez (1985).

The *Snowcrest trough* (Figure 2) lies west of the Gravelly Range in southwestern Montana and is evident on several thickness maps (Figures 2, 11, 12, 13, and 14).

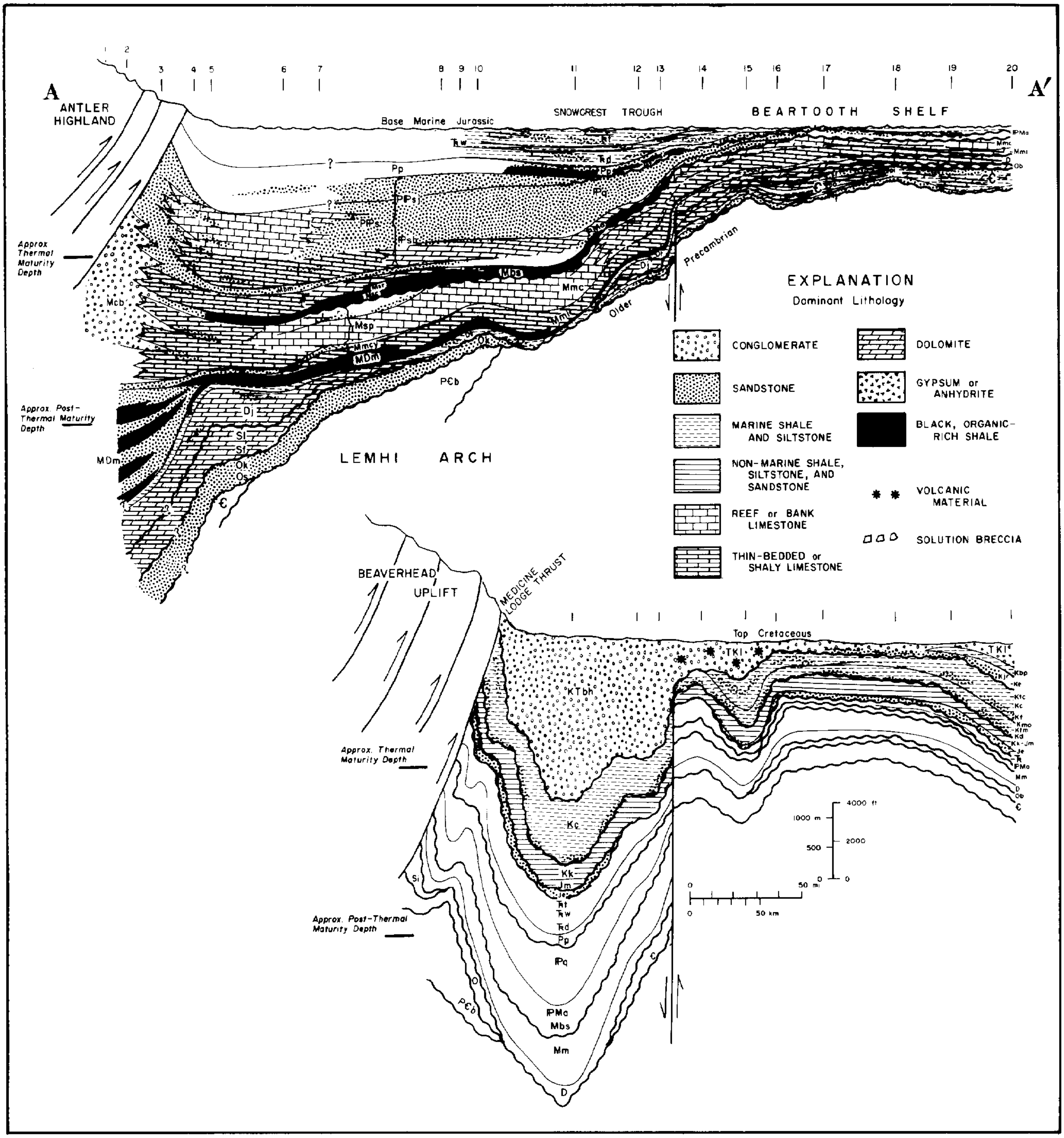

Figure 4—Southwest-northeast lithofacies cross section A-A′, from east-central Idaho to central Montana. Datum for upper section is the base of the Jurassic marine sedimentary rocks; datum for lower section is estimated sea level at the end of Cretaceous time. Section is palinspastically restored in thrust belt. Line of cross section is shown in Figures 2 and 3. Abbreviations for formation names used in Figures 4-7 are given on p. 63. Also given are the references for the numbered sections used in Figure 4-7.

TKl—Livingston Fm
KTbh—Beaverhead Fm
Tklh—Hoppers Fm
Tklb—Billman Cr. Fm
Tklmc—Miner Cr. Fm
TKLc—Cokedale Fm
TKs—Sedan Fm
TKlm—Maudlow Fm
TKw—Willow Cr. Fm
Klc—Landslide Fm
Kev—Everts Fm
Kem—Elkhorn Mts. Volcanics
Kgs—Golden Spike Cgl
Kcc—Carter Cr. Fm
Kj—Jens Fm
Kcb—Coberly Fm
Kbd—Dunkelberg Fm
Ksm—Slim Sam Fm
Kl—Lennep Ss
Kfg—Fox Hills Ss
Khc—Hell Cr. Fm
Kjr—Judith R. = River Fm
Kcl—Claggett Fm
Kn—Niobrara Fm
Kcl—Carlile Fm
Kg—Greenhorn Fm

Kmd—Muddy Fm
Kt—Thermopolis Sh
Ksc—Skull Cr. = Creek Fm
Kd—Dakota Fm
Ktc—Telegraph Cr. = Creek Fm
Kf—Frontier Fm
Kc—Colorado Sh
Kco—Cody Sh
Km—Montana Gp
Kmr—Marias R. Fm
Ksm—St. Mary R. Fm
Kh—Horsethief Ss.
Kbp—Bearpaw Sh
Ktm—Two Medicine Fm
Kvi—Virgelle Ss
Keu—Eagle Ss
Kmk—Kevin Sh
Kmf—Ferdig Mbr
Kmcc—Cone Calc. Mbr
Kb—Blackleaf Fm
Kbv—Vaughan Mbr
Kbt—Taft Hill Mbr
Kbf—Flood Mbr
Kk—Kootenai Fm
Jm—Morrison Fm
Je—Ellis Gp

Js—Swift Fm
Jr—Rierdon Fm
Jsa—Sawtooth Fm
Jp—Piper m
Trc—Chugwater Fm
Trt—Thaynes Fm
Trw—Woodside Fm
Trd—Dinwoody Fm
Pp—Phosphoria Fm
Pr—Retort Sh.
Pm—Meade Peak Sh
Pt—Tosi Chert
Ps—Shedhorn Ss
Pf—Franson Mbr
Pg—Grandeur Mbr
Pennsylvanian:
q—Quadrant Fm
a—Amsden Fm
Mbs—Big Snow Gp
Mh—Heath Fm
Mo—Otter Fm
Mk—Kibbey Fm
Mcb—Copper Basin Fm
Mmc—Mission Canyon Fm
Mml—Lodgepole Fm
Mc—Castle Reef Dol

Ma—Allan Mtn. Ls
Mcrs—Sun River Dol
Msr—Surrett Can. Fm
Msc—South Cr. Fm
Msp—Scott Peak Fm
Mmjc—Middle Canyon Fm
MDm—Milligen Can. Fm
Dt—Three Forks Fm
Dtp—Potlatch Mbr
Dj—Jefferson Fm
Db—Birdbear Fm
Dd—Duperow Fm
Dsr—Souris R. Fm
Dm—Maywood Fm
Sf—Fish Haven Fm
Sl—Laketown Fm
Ob—Big Horn Fm
Ok—Kinnikinnic Fm
Os—Summerhouse Fm
Csr—Snowy Range Fm
Cpi—Pilgrim Ls
Cpa—Park Sh
Cm—Meagher Ls
Cw—Wolsey Sh
Cf—Flathead Ss
PCb—Belt Supergroup

1. Central Pioneer Mtns., Idaho T. 4 N., R. 21, 22 E., Skipp et al., 1979b
2. Iron Bog Cr. area, T. 6 N., R. 23 E., Idaho, Skipp et al., 1979b
3. Cabin Cr. area, T. 6 N. - R. 22 E., Idaho, Skipp et al., 1979b
4. Timbered Dome area, T. 3 N. - R. 25 E., Idaho, Skipp et al., 1979a
5. Arco Hills, T. 4 N., R. 26, 27 E, Idaho, Skipp et al., 1979a
6. Howe Peak area, T. 4, 5 N., R. 28, 29 E., Idaho, Skipp et al., 1979a, b
7. East Canyon–Box Canyon, T. 6, 7 N., R. 29 E., Idaho, Skipp et al., 1979a, b
8. So. Beaverhead Mtns., T. 9 N. - R. 32 E., Idaho, Skipp et al., 1979a, b
9. Copper Mtn–Blue Dome area, T. 10 N., R. 30 E., Idaho, Skipp et al., 1979a; Scholten and Hait, 1962
10. Deadman Lake area, T. 16 S., R. 10, 11 W., Montana, Skipp et al., 1979b; Scholten and Hait, 1962
11. Lima Peaks–Red Peaks area, T. 15 S., R. 8, 9 W., Mont., Scholten, 1950; Moritz, 1960
12. Blacktail Cr., T. 12 S., R. 6 W. Mont., Keenmon, 1950
13. Snowcrest Range, T. 11, 12 S., R. 5 W., Mont., Gealy, 1953; Kummel, 1960; Cressman and Swanson, 1964
14. Gravelly Range, T. 10 S., R. 1, 2 W., Mont., Hadley, 1980; Mann, 1954; Cressman and Swanson, 1964
15. Sphinx Mtn. area., T. 8 S., R. 2 E., Mont., Beck, 1960; Gardner et al., 1946
16. Garnet Mtn. area, T. 6, 7 S., R. 4 E., Mont., McMannis and Chadwick, 1964; Schwartz, 1972
17. Hyalite Canyon, T. 4 S., R. 6 E., Mont., McMannis, 1962
18. Mill Cr., T. 6 S., R. 9, 10 E. Mont., Wilson, 1936; McMannis, 1962
19. Picket Pen Cr., T. 4 S., R. 14 E., Mont., Gardner et al., 1946
20. Continental, No. 1 Govt., Sec. 33, T. 2 S., R. 19 E., Mont., American Strat. Co.
21. NE Flint Cr. Range, T. 8, 9 N., R. 11 W., Mutch, 1960; Gwinn, 1965
22. Jefferson Canyon, T. 1, 2 N., R. 2 W., Mont, Alexander, 1955; Gardner et al., 1946
23. Logan Area, T. 2 N., R. 2 E., Mont., Robinson, 1963; Hanson, 1960; McMannis, 1962
24. Sixteen Mine–Maudlow area, T. 4, 5 N., R. 6, 7, 8 E., Mont., Gardner et al., 1946; Skipp and McGrew, 1977
25. SW Castle Mtns. area, T. 7, 8 N., R. 7, 8 E., Mont., Gardner et al., 1946; Tanner, 1949
26. Amerada Russell No. 1, Sec. 1, T. 8 N., R. 13 E., Mont, American Strat. Co.
27. Continental NP No. 1, T. 9 N., R. 17 E., Mont, American Strat. Co.
28. Cabinet Mtns., T. 22–29 N., R. 28–30 W., Mont., Aadland, 1979
29. Spotted Bear, T. 25 N., R. 14 W., Mont., Ross, 1959; Theodosis, 1955
30. Prairie Reef area, T. 23 N., R. 11 W., Mont., Mudge, 1972; Wilson, 1955; Theodosis, 1955
31. Sun River Can., T. 22, 23 N., R. 9, 10 W., Mont., Mudge, 1972; Wilson, 1955
32. Phillips, Yeager-1, Sec. 6, T. 23 N., R. 6 W., Mont., American Strat. Co.
33. Brit. Amer., Severson-1, Sec. 7, 24 N., 2 W., Mont., American Strat. Co.
34. Gen. Pet., Holt-1, Sec. 30, 25 N, 1W., Mont., American Strat. Co.
35. Continental-1, State, 25 N., 1 E., Mont, American Strat. Co.
36. Huntley-1, Rossmiller, 27-26 N, 3 E., Mont., American Strat. Co.
37. Chamberlain, 1-Higgins, 30–27 N, 6E., Mont., American Strat. Co.
38. Texas Co., Kiemele-1, 26–31 N, 13 E., Mont., American Strat. Co.
39. So. Union, N. Chinook-1, 16–35 N., 19E., Mont., American Strat. Co.
40. Feather Woman Mtn. area, T. 29 N., R. 10, 11, W., Mont, Wilson, 1955
31. See Figure 6
32. See Figure 6
41. Phillips-1, Randall, 6–21 N., 5 W., Mont., American Strat. Co.
42. Schoonmaker-1, Stephan, 12–20N, 1E., Mont.
43. Riverdale, Murphy-1, 8–17N., 2E., Mont., American Strat. Co.
44. Smith R. Canyon, 13, 14, 15 N., 3, 4 E., Mont., Walker, 1974; Dahl, 1971
45. White Sulphur Spgs. area, 9, 10 N., 7E., Mont., Tanner, 1949; Dahl, 1971; Billings Geol. Soc., 1962
25. See Figure 5
24. See Figure 5
46. N. Bridger Mts., 3, 4 N., 6 E., Mont., Klemme, 1949; McMannis, 1955; Skipp and McGrew, 1977
47. Livingston area, 2, 3S., 9E., Mont., Gardner et al., 1946; Richards, 1957; Sando, 1972; Brown, 1957
18. See Figure 4
48. Cinnabar Mtn., 8S, 8E., Mont., Wilson, 1934; Kummel, 1960; McMannis, 1962
49. NW Yellowstone Park, Wyo. and Mont., Ruppel, 1972; Brown, 1957; McMannis, 1962

Figure 4 Legend

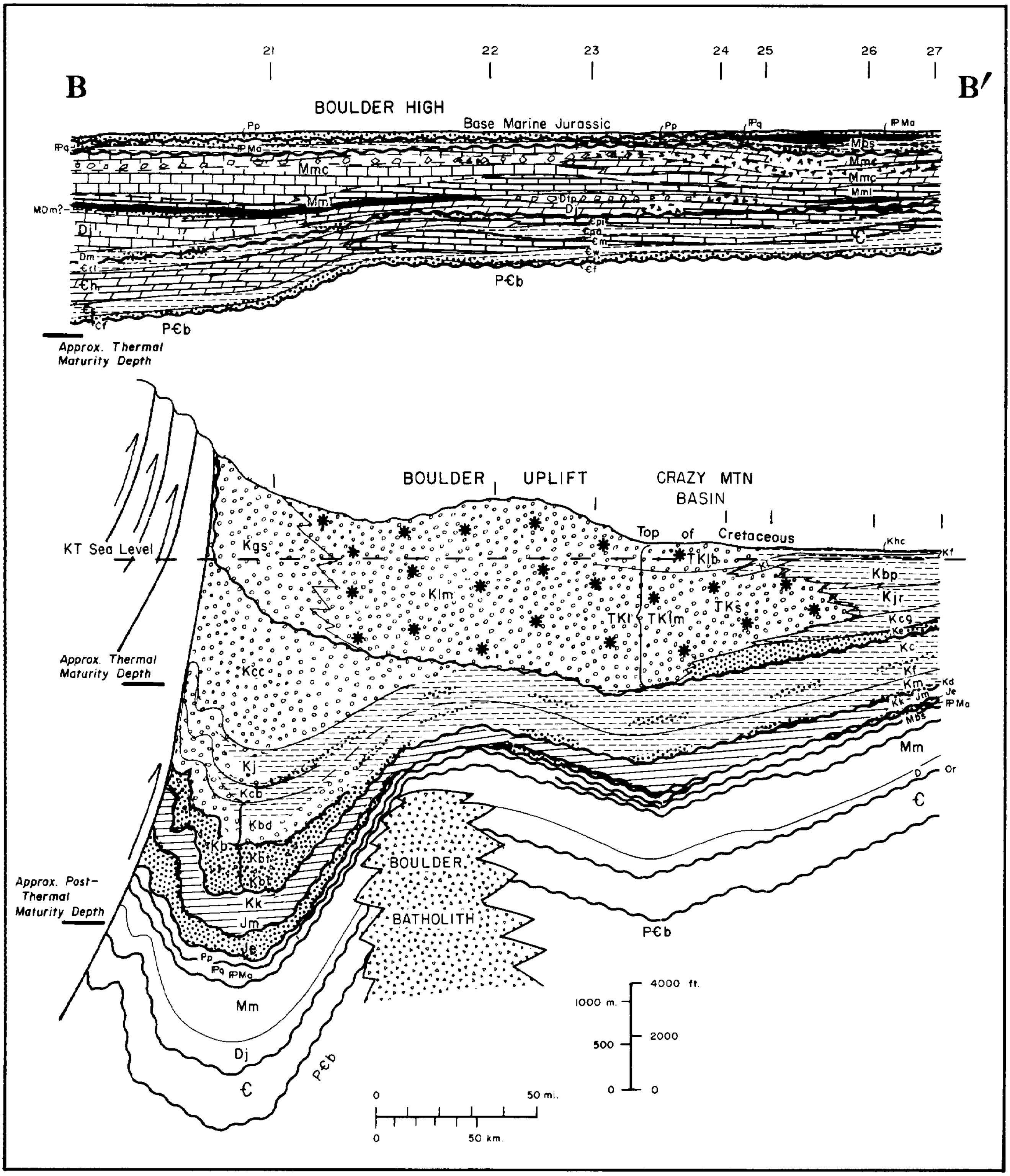

Figure 5—West-east lithofacies cross section B-B′, from west-central to east-central Montana. Datum for upper section is the base of the Jurassic marine sedimentary rocks; datum for lower section is estimated sea level at the end of Cretaceous time. Section is palinspastically restored in thrust belt. Line of cross section is shown in Figures 2 and 3. Lithologic explanation is shown in Figure 4, and explanation of formation abbreviations and numbered sections is given in Figure 4 legend.

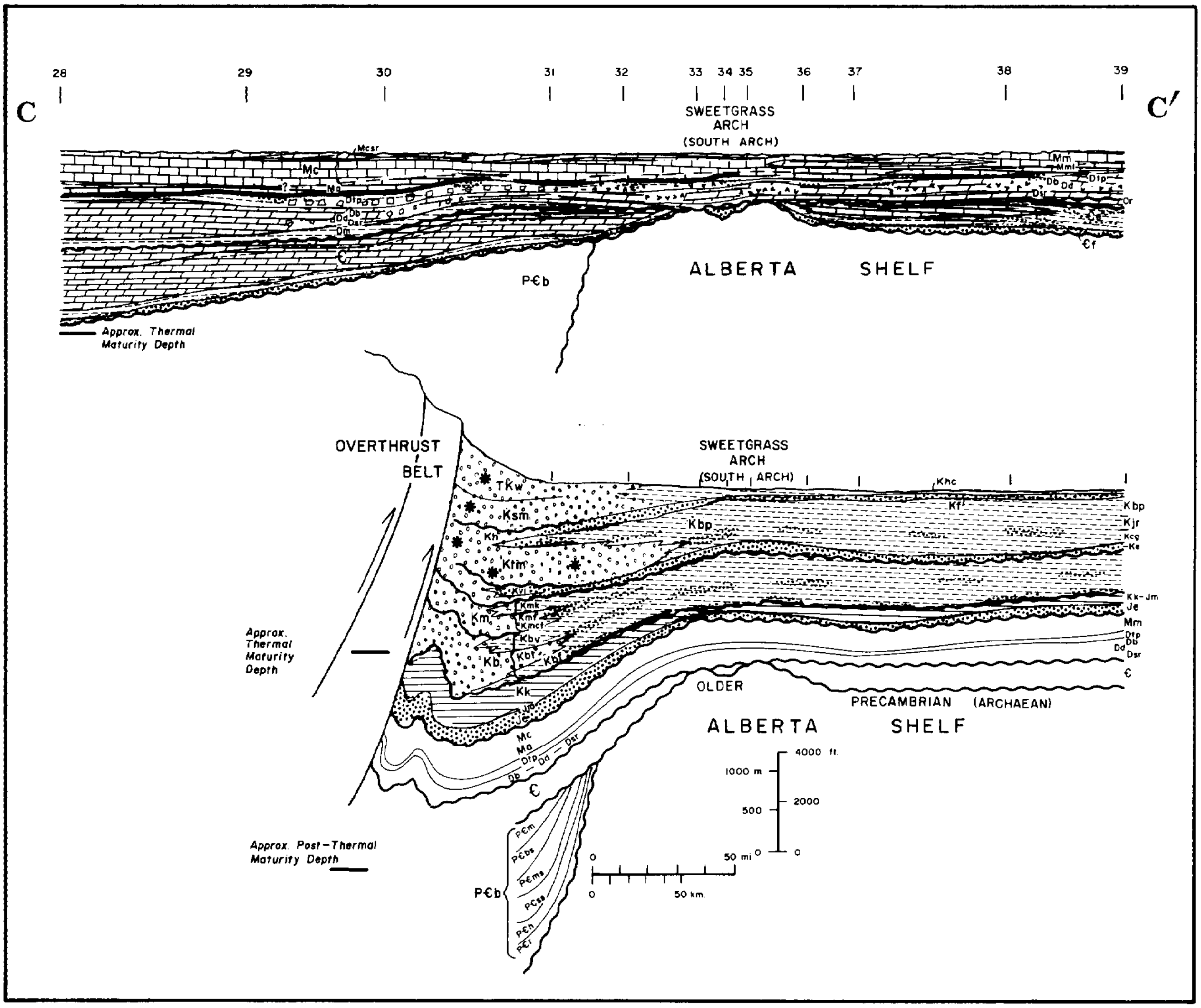

Figure 6—West-east lithofacies cross section C-C′, from northwestern Montana to north-central Montana. Datum for upper section is the base of the Jurassic marine sedimentary rocks; datum for lower section is estimated sea level at the end of Cretaceous time. Section is palinspastically restored in thrust belt. Line of cross section is shown in Figures 2 and 3; Lithologic explanation is shown in Figure 4, and explanation of formation abbreviations and numbered sections is given on Figure 4 legend.

STRATIGRAPHY AND SEDIMENTARY FACIES

Precambrian

Metamorphosed older Precambrian rocks are exposed in several mountain ranges north and west of Yellowstone National Park. These rocks have been studied by several authors, including Reid (1957), Foose et al. (1961), Stewart (1972), James and Hedge (1980), and others.

Proterozoic rocks of the Belt Supergroup make up most of the pre-Tertiary exposures in northwestern Montana and are present in some mountain ranges east of the overthrust belt. They are greater than 50,000 ft (15,000 m) thick in the vicinity of the Coeur D'Alene trough (Figure 8) and thin to zero roughly at the leading edge of the thrust belt in northwestern and southwestern Montana and along the "Perry line" near Bozeman, Montana (Figures 3, 4, 5, and 6). Belt equivalent rocks have been identified in boreholes east of Bozeman in the central Montana trough, or "Belt embayment," area at least as far east as Billings. As pointed out by Harrison et al. (1974), much of the paleostructural grain of western Montana was developed at least as early as late Precambrian time and most of it was retained through most of the remainder of geologic time. Among the more persistent of the older features are the Coeur D'Alene trough, Lemhi arch, central Montana trough, Beartooth shelf, Alberta shelf, and perhaps the Boulder high or ancestral Belt Island (Jurassic) complex. Much of the Belt Supergroup sequence is composed of relatively fine-grained clastic rocks, although a narrow band of boulder conglomerate debris, the Lahood facies (McMannis, 1963), is present along the northwest border of the Beartooth shelf, and some coarse sandstone and pebbly beds occur in southwesternmost Montana in the vicinity of the Lemhi arch. These beds were apparently deposited along a fault-controlled shoreline in the vicinity of the present-day Horse Prairie fault (D. A. Lopez, personal communication, 1984).

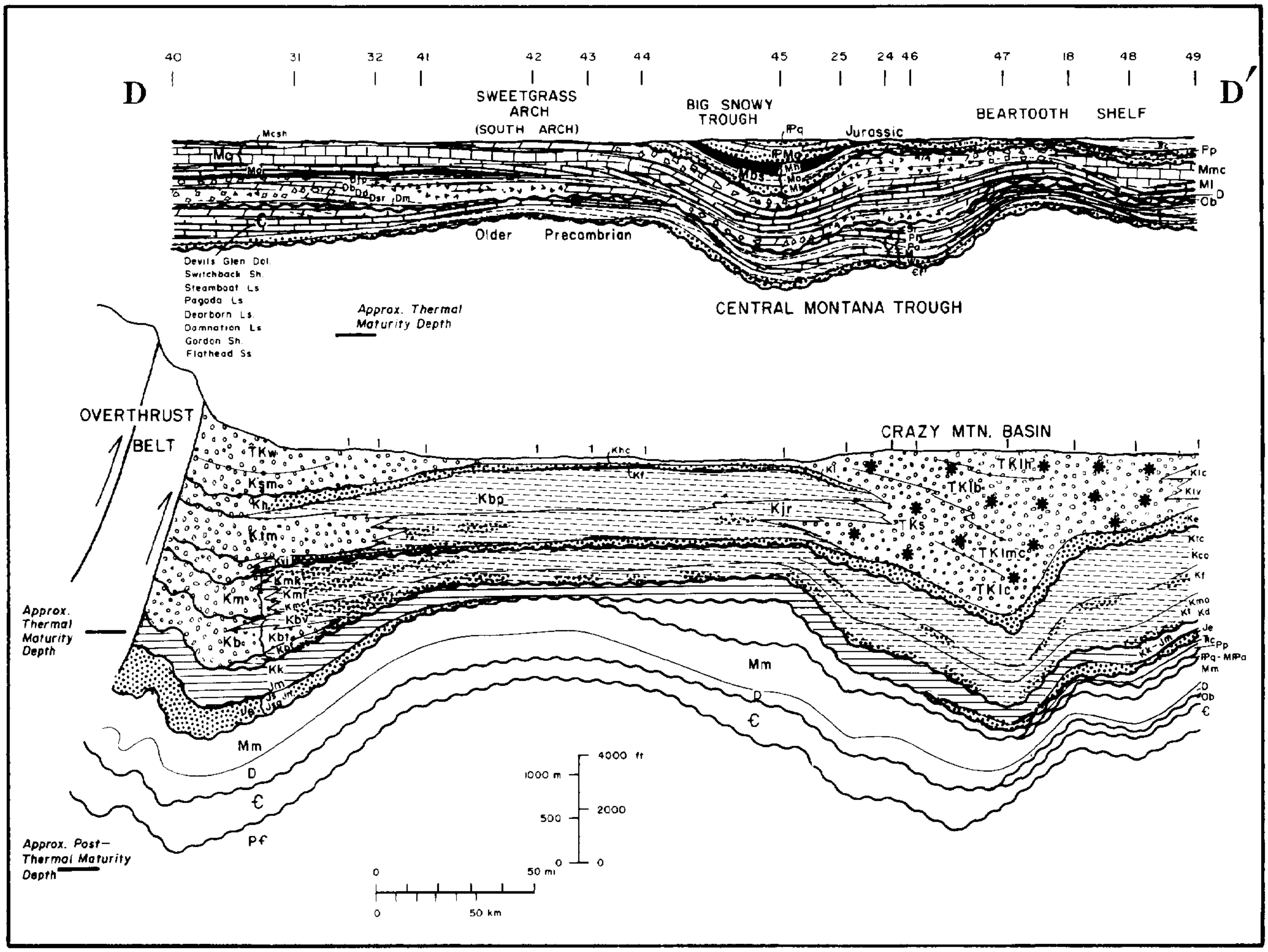

Figure 7—North-south lithofacies cross section D-D′, from south of Glacier National Park to Yellowstone National Park. Datum for upper section is the base of the Jurassic marine sedimentary rocks; datum for lower section is estimated sea level at the end of Cretaceous time. Section is palinspastically restored in thrust belt. Line of cross section is shown in Figures 2 and 3. Lithologic explanation is shown in Figure 4, and explanation of formation abbreviations and numbered sections is given in Figure 4 legend.

A broad band of thick stromatolitic algal carbonate facies (Helena Formation and equivalents) is present on the west flank of the Alberta shelf and extends from near the Canadian border southward at least as far as the Helena vicinity (Figure 8; Theodosis, 1955; McGill and Sommers, 1967).

Cambrian

A transgressive sequence of Middle and Upper Cambrian rocks over 1000 ft (300 m) thick is present in west-central Montana and thickens to more than 2000 ft (600 m) in scattered exposures west of Butte (Figures 5, 6, and 9). The main trend of thickening follows the Coeur D'Alene–central Montana trough. The section is less than 150 m thick in the vicinity of the Sweetgrass arch area and thins to zero in southwestern Montana where a coarser clastic facies occurs near the Lemhi arch source area. Cambrian rocks are somewhat thicker in the area of the Snowcrest trough (Figure 4).

The main clastic source areas during Cambrian time were located in the continental interior on the Canadian shield and the Transcontinental arch (Figure 1). The widely transgressive Flathead Sandstone is present at the base of the sequence almost everywhere and ranges in age from Middle to Upper Cambrian. In the Williston basin of eastern Montana and North Dakota, Cambrian rocks comprise a sandstone and shale facies that grades westward to green shale and limestone facies in central Montana. Approximately in the position of the central and northern overthrust belt, the Cambrian sequence becomes dominated by a thick section of sucrosic dolomite (Hasmark Formation and equivalents) (Hanson, 1952), which is more than 2000 ft (600 m) thick west of Helena and Butte, Montana (Figures 5 and 9). The dolomite facies is generally unfossiliferous, but careful outcrop and microscopic studies reveal evidence of bioclastic, algal, and oolitic fabric that probably originated as carbonate bank buildups in the boundary belt between the early Paleozoic shelf and miogeosyncline.

Ordovician-Silurian

Ordovician rocks are absent in most of western Montana except for the Beartooth shelf area where the Bighorn Dolomite, as thick as 300 ft (90 m), is present. The zero

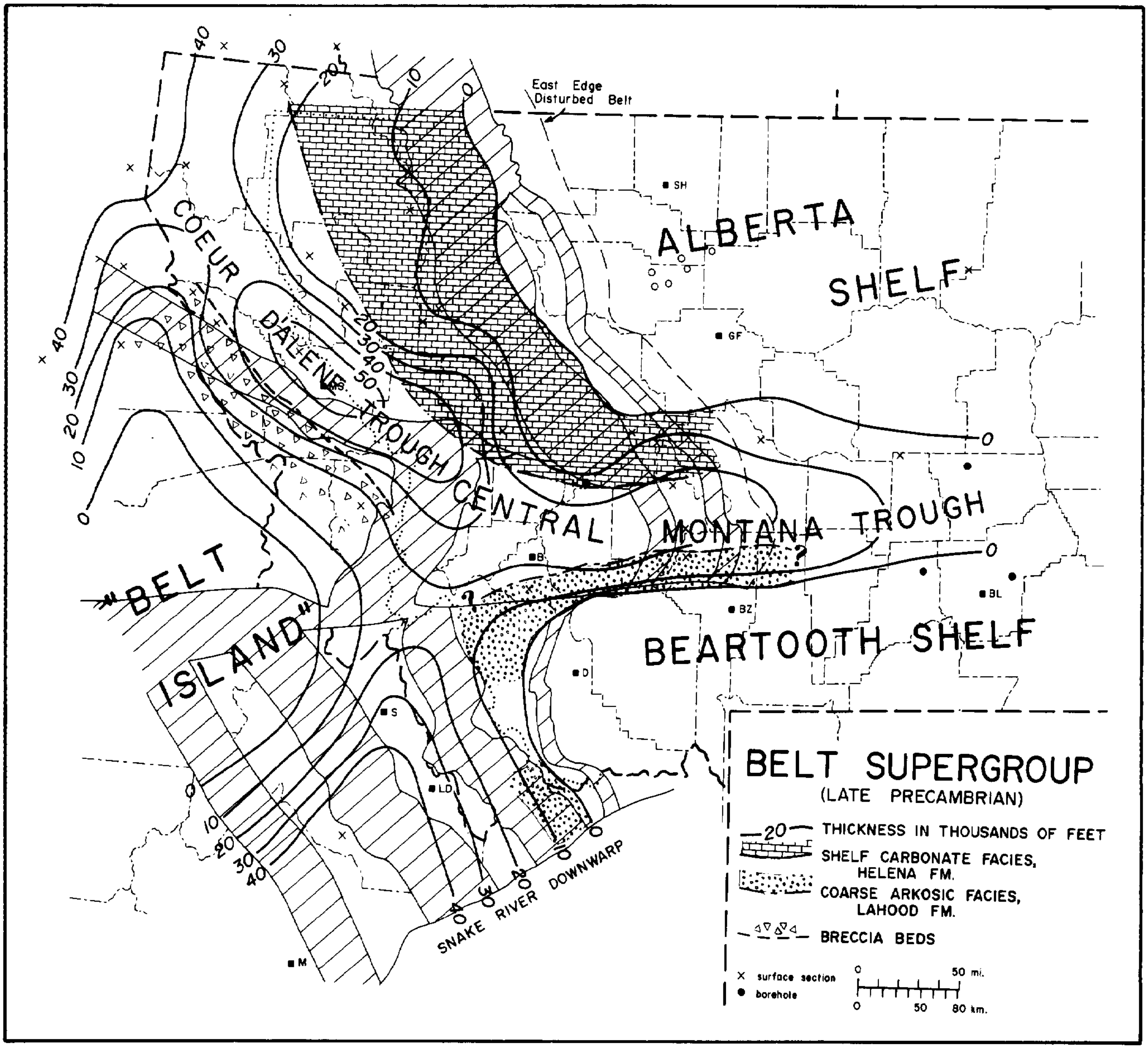

Figure 8—Thickness and general lithofacies of the upper Precambrian Belt Supergroup, partly restored. Data are palinspastically restored in thrust belt. Modified in part after Harrison (1972).

edge of the Bighorn extends along the north and west borders of the Beartooth shelf, projects slightly eastward into the area of the central Montana trough, and continues northward across the Alberta shelf into Canada (Figure 10). To the east, Ordovician rocks thicken uniformly into the Williston basin, where they reach thicknesses of more than 1000 ft (300 m). Ordovician rocks are very thick in the Muldoon trough of east-central Idaho (Kinnikinnic Quartzite) but they wedge out rapidly eastward and are truncated by Upper Devonian erosion approximately along the Montana–Idaho border (Figures 4 and 10) (James and Oaks, 1977; Scholten, 1967).

Silurian rocks have not been identified in western Montana, although the thick section of Silurian dolomite (Laketown Formation) in east-central Idaho pinches out eastward very close to the border between southwestern Montana and east-central Idaho a short distance west of the truncated edge of Ordovician rocks in the vicinity of the Lemhi arch (Figures 4 and 10). This erosional offlap relationship between the Bighorn–Interlake and the Kinnikinnic–Laketown formations and the overlying Upper Devonian beds, plus the general absence of sandstone and siltstone in the Silurian and Ordovician carbonate beds, suggest that the erosional edges of both units are some distance from the Ordovician or Silurian shorelines and that a broad area of originally deposited carbonate facies was removed by pre-Middle or Upper Devonian erosion.

Devonian

During Late Silurian through Middle Devonian time, most of the Rocky Mountain shelf was apparently emergent, and Lower and Middle Devonian rocks are absent over all of Montana except for scattered deposits in southern Montana

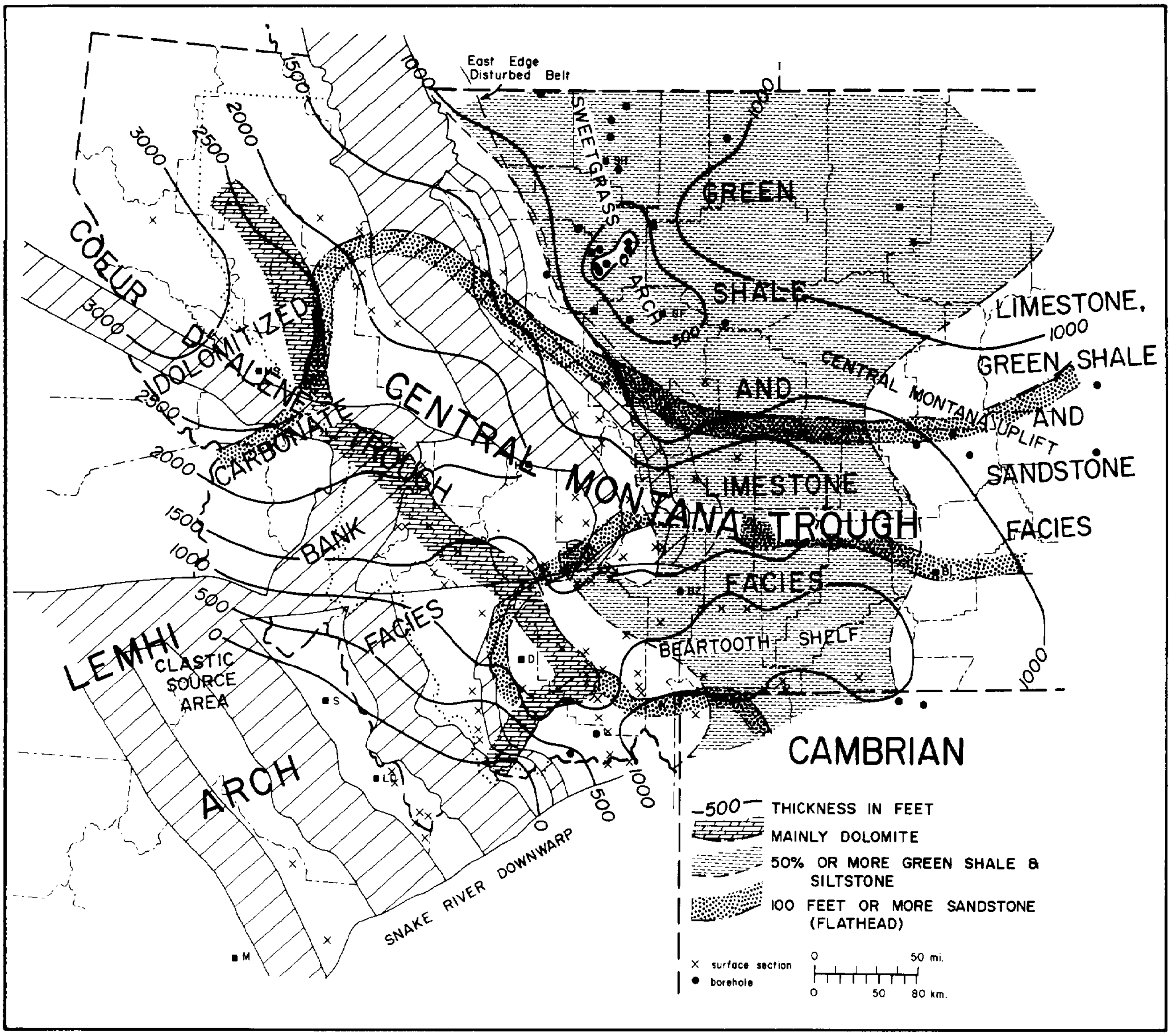

Figure 9—Thickness and general lithofacies of Cambrian rocks, partly restored. Data are palinspastically restored in thrust belt.

and a section that is increasingly older eastward into the Williston basin. Widespread but gentle erosion of the Silurian and Ordovician beds occurred at this time and resulted in the regional truncation offlap relationship between Upper Devonian rocks and the underlying units (Figure 4).

The worldwide Upper Devonian transgression appears to have spread completely across the Rocky Mountain shelf and covered all of Montana. Upper Devonian rocks comprise a transgressive sequence of fine- to medium-grained, red to green, clastic marine beds (Maywood or lower Souris River formations) overlain by a prominent sequence of generally porous stromatoporoid and coral-bearing dolomite, fossiliferous limestone, evaporite beds, and shaly limestone (Jefferson Formation) that form the bulk of the Devonian section. The strongly cyclic nature of this sequence has been described by several authors (Wilson, 1955; Sandberg and Hammond, 1958; Rose, 1976). The Jefferson is overlain by regressive green and red shale, siltstone, evaporite beds, and shaly limestone beds of the Three Forks Formation, which grade upward into the Upper Devonian-Mississippian dark organic shale and siltstone beds of the Bakken and Sappington formations. Lithologic aspects and the facies distribution of these beds have been described in detail by Rau (1962), McMannis (1962), and Gutschick (1962).

The Upper Devonian rock sequence is more than 1000 ft (300 m) thick in most of western Montana and more than 2000 ft (600 m) thick in much of the western thrust belt area. It thickens rapidly westward into the Muldoon trough along the southwestern Montana-Idaho border (Figures 4 and 10). The western facies is dominated by dolomitized stromatoporoid and tabulate coral-bearing carbonate bank deposits commonly containing thick beds of highly porous sucrosic dolomite. The continued influence of the central

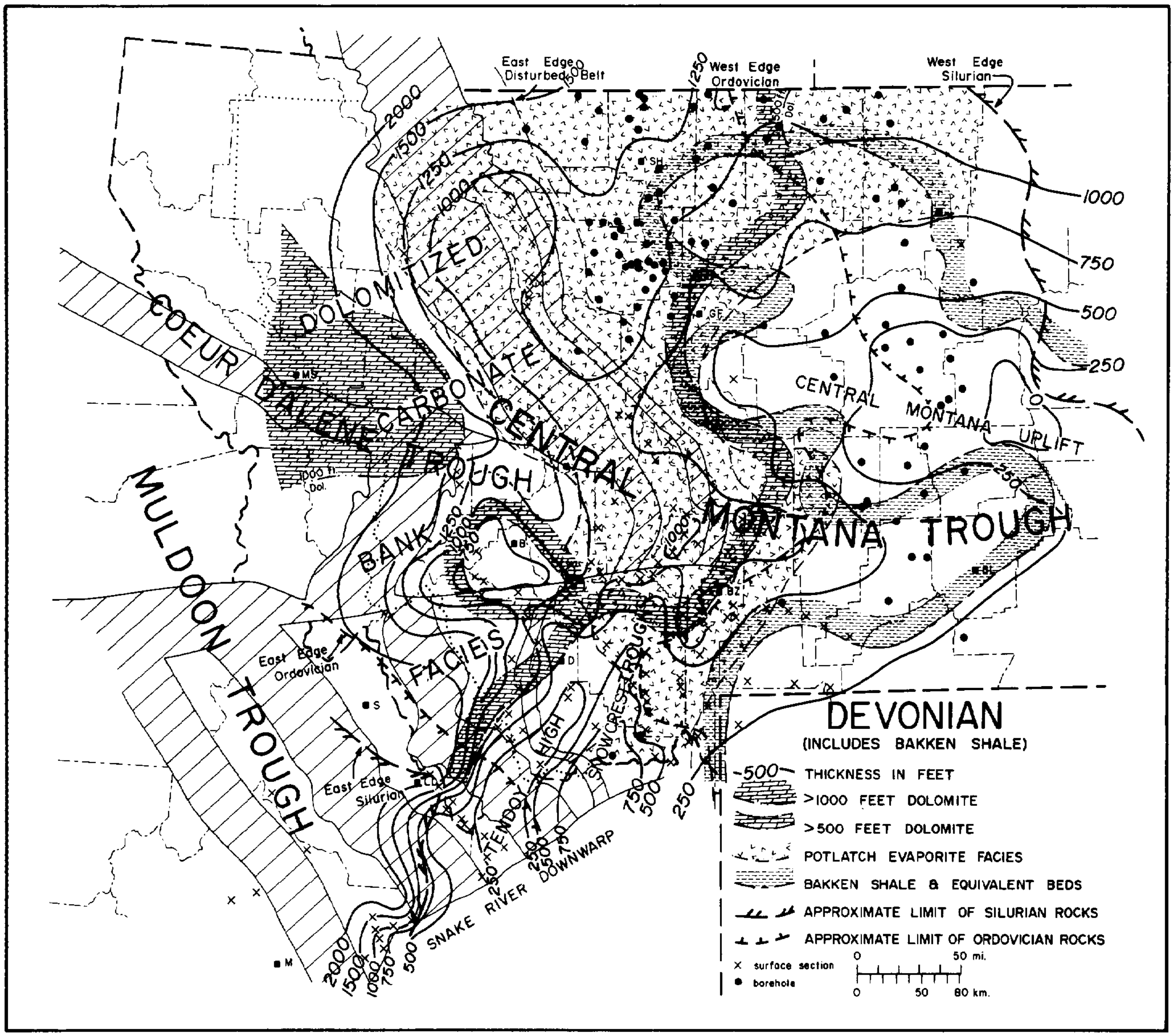

Figure 10—Thickness and general lithofacies of Devonian rocks, partly restored. Data are palinspastically restored in thrust belt. Approximate erosional limits of Silurian and Ordovician rocks are shown.

Montana trough is evident from the thick west-east trend in about the same position as that of the late Precambrian and Cambrian intervals. A relatively prominent area of thinning is present in the vicinity of the Lemhi arch in southwestern Montana ("Tendoy high"), and an accompanying thick southwest-northeast trend is present just to the east in the Snowcrest trough. The relatively prominent thin northwest-southeast area that coincides closely with the northwestern overthrust belt (Figure 10) may be related to outcrop leaching of gypsum or anhydrite beds of the Potlatch facies and similar beds in the underlying Jefferson Formation. Marked thinning or absence of Upper Devonian beds occurs along the central Montana uplift and extends westward along the Little Belt Mountains area south of Great Falls. The western part of this belt may also be related to evaporite solution of outcrop sections, although the absence of lower Three Forks Potlatch breccias in the Little Belt Mountains suggests otherwise. The accompanying absence of the Devonian-Mississippian Bakken Formation beds in this area suggests that broad, gentle uplift occurred here at this time. The distribution of the Bakken and equivalent facies in south central Montana also demonstrates the continued influence of the central Montana trough and the Beartooth shelf (Figure 10).

The regional distribution of evaporite beds and related solution breccias in the Jefferson and Three Forks formations suggests that evaporitic conditions occurred in a back bank cabonate environment immediately east of the north-south-trending belt of carbonate bank facies buildup in westernmost Montana (Figure 10). This paleogeographic relationship may have resulted in the establishment of a westward-directed hydrodynamic gradient that enabled high

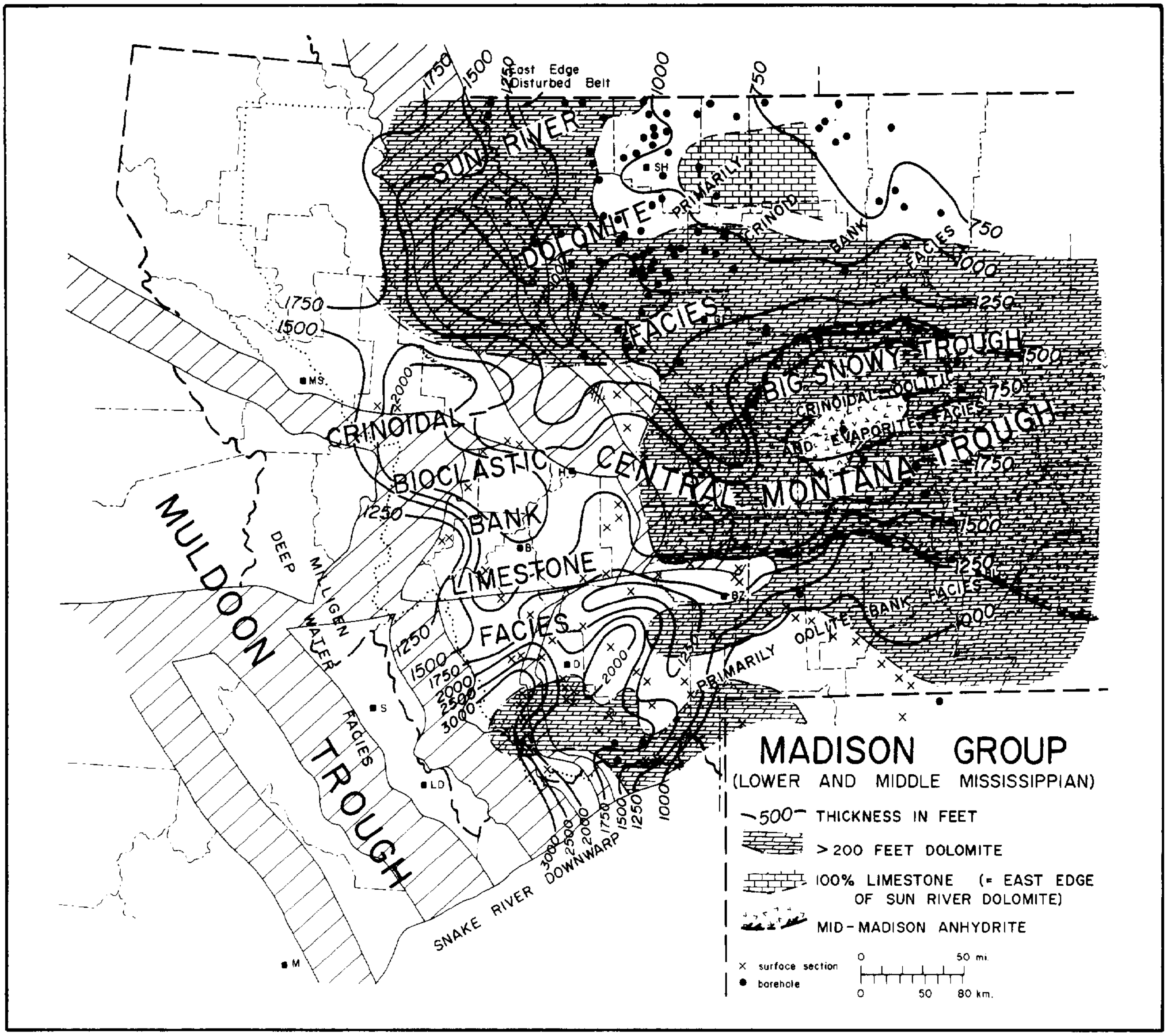

Figure 11—Thickness and general lithofacies of Lower and Middle Mississippian Madison Group, partly restored. Data are palinspastically restored in thrust belt.

magnesium waters of the shelf environment to move westward during low sea level evaporitic stages of the cycles, which caused early dolomitization of the thick carbonate facies by seepage reflux processes. Such a mechanism may have been in effect during much of the early to middle Paleozoic, as suggested by the similarity in gross dolomitic facies distribution of the Cambrian sequence. Evidence for shelf back bank evaporitic environments, however, is lacking in the earlier Paleozoic sequences.

Detailed lithologic and stratigraphic aspects of Devonian rocks in western Montana are described by Sloss and Laird (1945, 1946, 1947), Wilson (1955), Rau (1962), Sandberg (1962, 1965), Sandberg and McMannis (1964), Benson (1966), and Sandberg and Mapel (1967).

Lower-Middle Mississippian: Madison Group

The paleogeographic setting characteristic of the Devonian persisted into the Mississippian, and cyclic carbonate deposition predominated throughout the Lower and Middle Mississippian. These carbonate beds are not nearly as highly dolomitized as those of the Devonian, and evaporite deposits are not as widespread in the shelf area. Partly for these reasons, cycles in Mississippian rocks are more difficult to recognize. A thick crinoidal, bioclastic bank facies is present in approximately the same position as the thick Devonian stromatoporoid, coral bank facies (Figure 11). Evaporite beds east of the bank facies, however, are generally restricted to the central Montana trough where a prominent anhydrite or gypsum unit (middle Madison in age) is widely recognized in the subsurface and is represented by a prominent breccia interval in outcrop sections ("lower solution zone" of Sando, 1972, 1976). A thick trend of dolomite and anhydrite facies is generally prevalent in the area of the central Montana trough, which indicates continued relatively greater subsidence in that area. Thinning of the Madison section occurs on the Alberta shelf

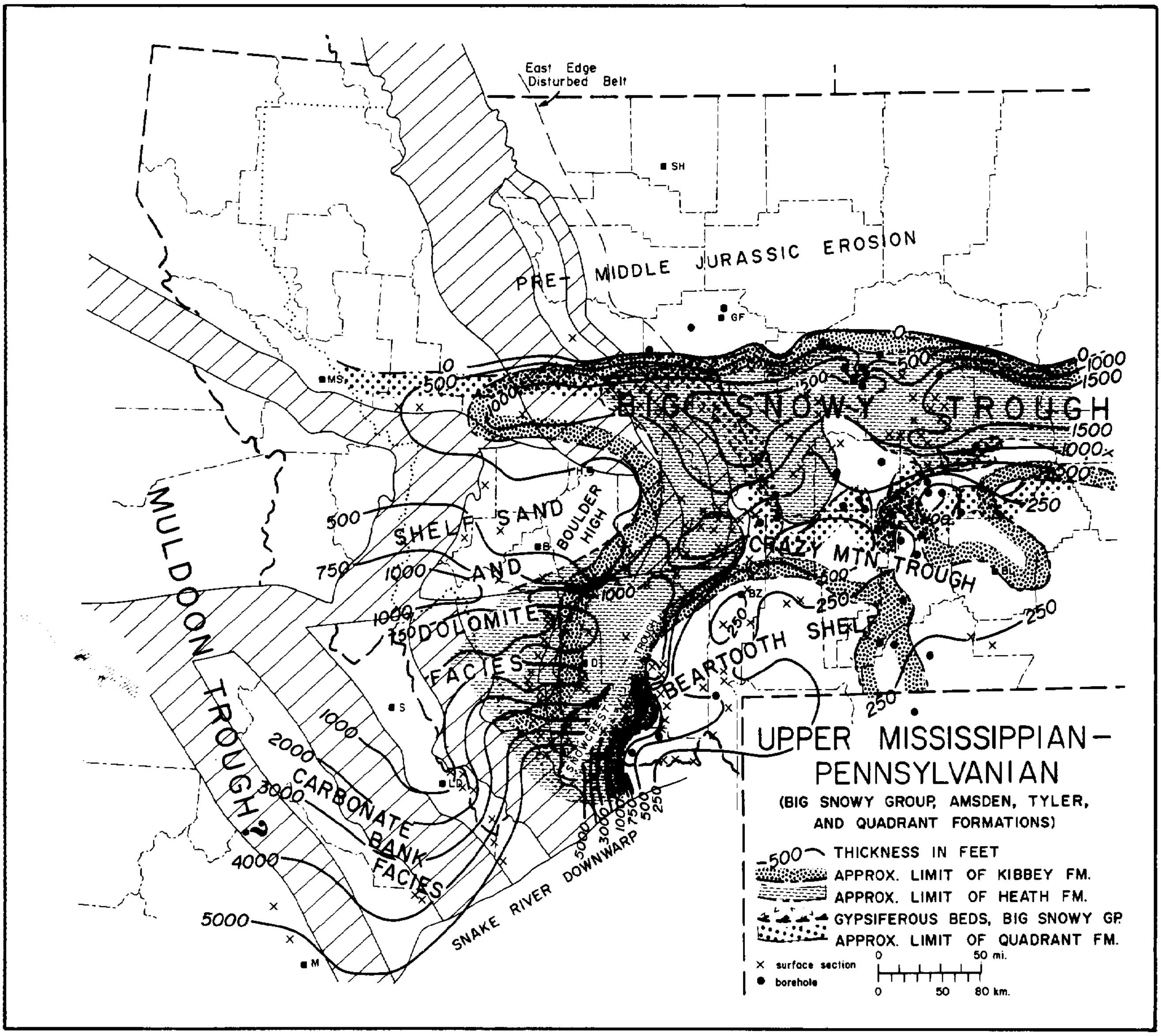

Figure 12—Thickness and general lithofacies of Upper Mississippian and Pennsylvanian rocks, partly restored. Data are palinspastically restored in thrust belt.

(Figure 11), which is characterized by crinoidal bank limestone facies. Thinning also occurs in the Beartooth shelf area south of the evaporitic trough belt where the carbonate beds are dominated by oolitic bank limestone facies.

The thick Madison carbonate bank facies in western Montana contains much less dolomite than the bank facies of the Devonian, which is almost entirely dolomite. The reasons for this may be related partly to generally diminished evaporitic conditions on the shelf during Madison deposition, which would have decreased the potential for development of westward-gradient seepage reflux systems. An additional reason may be that the Madison crinoid and oolite bank sediments contained considerably less original porosity than did the stromatoporoid and coral reefoid bank Devonian facies.

Detailed descriptions and analyses of the Madison rocks in the northern Rocky Mountains are published by several authors, including Sloss and Hamblin (1942), Sloss and Laird (1945), Nordquist (1953), Andrichuk (1955), Mudge et al. (1962), Roberts (1966, 1979), Huh (1967), Sando (1967, 1972, 1976), Sando et al. (1969), Smith (1972), Rose (1976), Skipp et al. (1979), Smith and Gilmour (1979), and Peterson (1984b).

Upper Mississippian–Pennsylvanian

Upper Mississippian and Pennsylvanian beds (Big Snowy Group, Amsden and Quadrant Formations) are dominated by clastic sediments and represent a marked change in regional depositional and tectonic conditions of the western North American continental shelf at this time. These beds contrast sharply with the underlying stable shelf carbonate sediments that dominated the Rocky Mountain shelf and the adjacent border of the Cordilleran miogeosyncline during Ordovician, Silurian, Devonian, and Early–Middle Mississippian time. Tectonic activity in the middle and southern Rocky Mountains area that began in Late

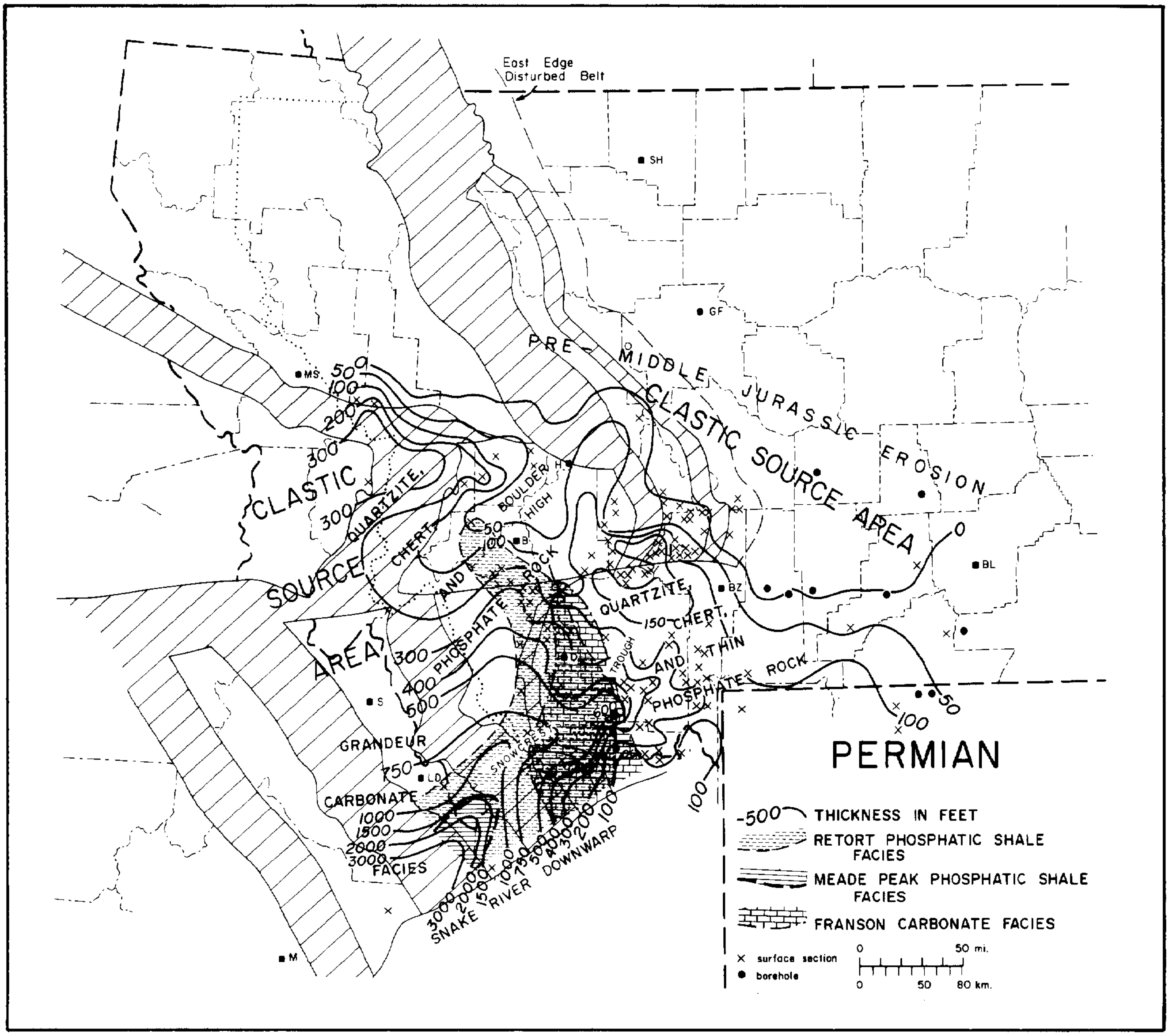

Figure 13—Thickness and general lithofacies of Permian rocks, partly restored. Data are palinspastically restored in thrust belt.

Mississippian time resulted in active growth of several prominent local uplifts and accompanying basins (Ancestral Rocky Mountains) and the rise of many new clastic source terranes. At the same time, the supply of clastic material from the Canadian Shield and Transcontinental arch increased and spread across the northern Rocky Mountain shelf. The regional paleogeographic and paleotectonic patterns in Montana appear to have remained essentially the same, but carbonate deposition was greatly diminished because of the increased supply of clastic material into the shallow water marine environment. Most of the organic carbonate bank growth that was characteristic of the middle Paleozoic time was eliminated on the shelf or craton edge during Late Mississippian–Pennsylvanian time. Farther west, however, at greater distances from clastic source areas along the western border of the shelf and the eastern flank of the Muldoon trough in east-central Idaho, a prominent carbonate bank facies is present (Figures 4, 11, and 12).

A substantial amount of the Upper Mississippian–Pennsylvanian section was removed by pre-Middle Jurassic erosion, but in general the major paleostructural elements of central to western Montana are reflected by the thickness and facies distribution patterns of this interval, particularly in the Upper Mississippian Big Snowy Group. Marked thickening of the section continues in the area of the central Montana trough, although this feature appears to separate into two west–east trough elements, the Crazy Mountain trough and the more sharply defined Big Snowy trough where the Big Snowy beds are more than 1500 ft (450 m) thick (Figures 7 and 12) and contain the thickest facies of the highly organic-rich Heath Formation. This thickness pattern is accompanied by more complex sedimentary facies relationships, and it is thought to reflect increased tectonic activity along the southern border of the Alberta shelf (Fanshawe, 1978; Maughan and Roberts, 1967). The Crazy Mountain trough merges westward with the north extension

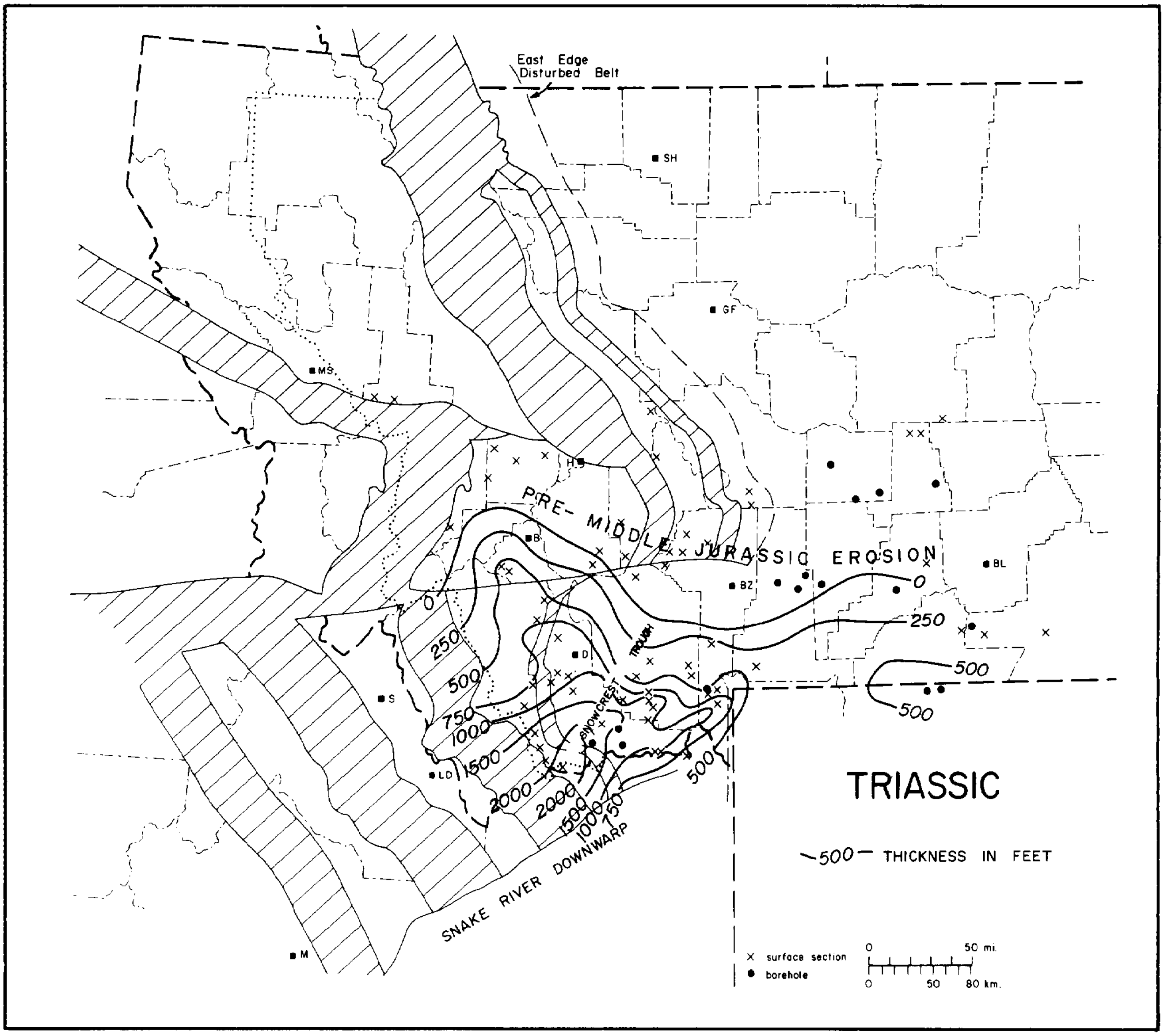

Figure 14—Thickness of Triassic rocks, partly restored. Data are palinspastically restored in thrust belt.

of the Snowcrest trough, which contains a substantially thicker section of Upper Mississippian and Pennsylvanian beds and tends to confine the southern extension of the Big Snowy facies. A markedly thinner sequence is present on the Beartooth shelf, and the Upper Mississippian–Pennsylvanian section is absent on the Alberta shelf. Thinning in both these areas is partly caused by pre-Middle Jurassic erosion, but they probably were also mildly elevated submarine shelf areas at this time. Restored thickness patterns show probable thinning in the vicinity of the Boulder high and the general area of the Lemhi arch (Figure 12).

A steadily increasing supply of quartzose sand from the northern source terrane is demonstrated by the almost complete dominance of clean shallow water marine and eolian sandstone beds in the upper part of the Upper Mississippian–Pennsylvanian clastic sequence (Quadrant Formation) (Figures 4, 5, 6, 7, and 12). This facies intertongues southwestward with dolomite and limestone beds of the carbonate bank facies along the east flank of the Muldoon trough and adjacent shelf margin. Much of the thick quartzose sand section is quartzitic or dolomitic, but substantial parts of it are composed of clean, well-sorted, porous sandstone in parts of southwestern Montana (D. Saperstone, personal communication, 1984).

Detailed studies of Upper Mississippian–Pennsylvanian rocks have been published by Sloss and Laird (1946), Easton (1962), Maughan and Roberts (1967), Harris (1972), Breuninger (1976), Rose (1976), Skipp et al. (1979), Smith and Gilmour (1979), and Maughan (1984).

Permian

The Permian section has a truncation offlap relationship with the underlying Pennsylvanian beds as the result of pre-Middle Jurassic erosion. The Permian sedimentary facies in southwestern Montana is characterized by phosphate, carbonate, and chert beds that represent the northern facies

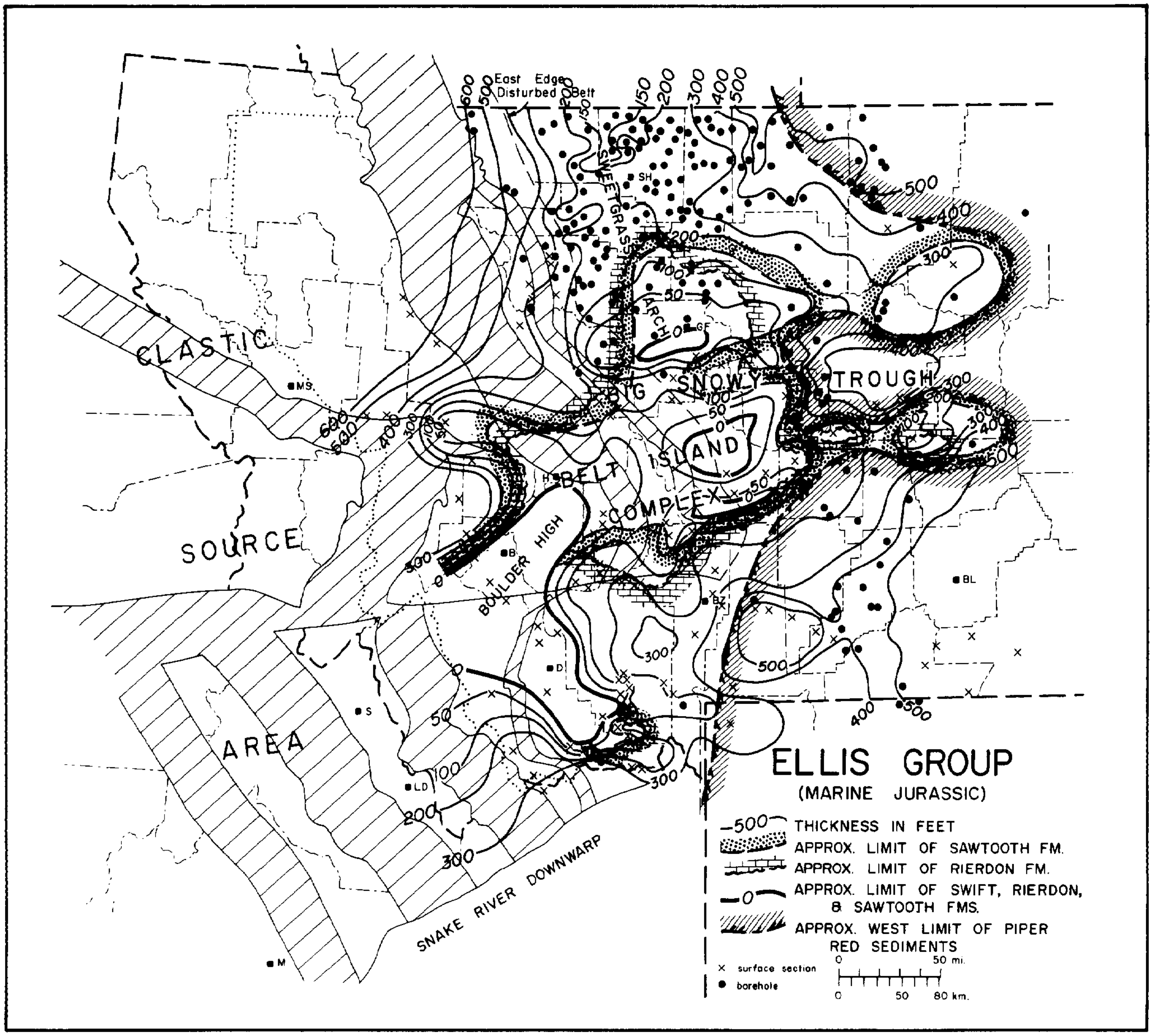

Figure 15—Thickness of the Jurassic marine Ellis Group, partly restored. Data are palinspastically restored in thrust belt.

of the Phosphoria basin of southeastern Idaho and western Wyoming. The northern clastic source area remained active during this time, as demonstrated by the presence of a quartzitic facies of Late Permian age (Shedhorn Sandstone) bordering the area of truncation (Figures 7 and 13). Evidence for a rejuvenated clastic source area in the vicinity of the Paleozoic Lemhi arch (Figures 2 and 13) is indicated by increased quartzose sand content of Permian beds west and northwest of Dillon. In southwesternmost Montana and adjacent Idaho, the Lower Permian section (Wolfcampian) contains a great thickness of carbonate beds (Grandeur Member), which represent an uninterrupted continuation of the depositional environments of Late Pennsylvanian time in that area (Figure 4). The major change in environmental conditions occurred at the close of the Grandeur cycle with the deposition of the overlying phosphate (Meade Peak and Retort), chert (Tosi), and carbonate (Franson) beds of the Phosphoria Formation facies, which dominate the Permian section in southwestern Montana. The thickest post-Grandeur Permian deposits occur in the Snowcrest trough and in the adjacent area of subsidence immediately to the west. This belt falls generally between the western and eastern quartzitic facies and south of the Boulder high (Figure 13).

Because of the presence of economic phosphorite deposits, the Permian beds have been thoroughly studied. Many publications are available that contain details of the stratigraphy, facies distribution, depositional environments and biostratigrapahy of this section in the western Rocky Mountain region (Cressman, 1955; McKelvey et al., 1959; Cressman and Swanson, 1964; Sheldon, 1963; Yochelson, 1968; Peterson, 1972a, b, 1980, 1984a, b; Maughan, 1975).

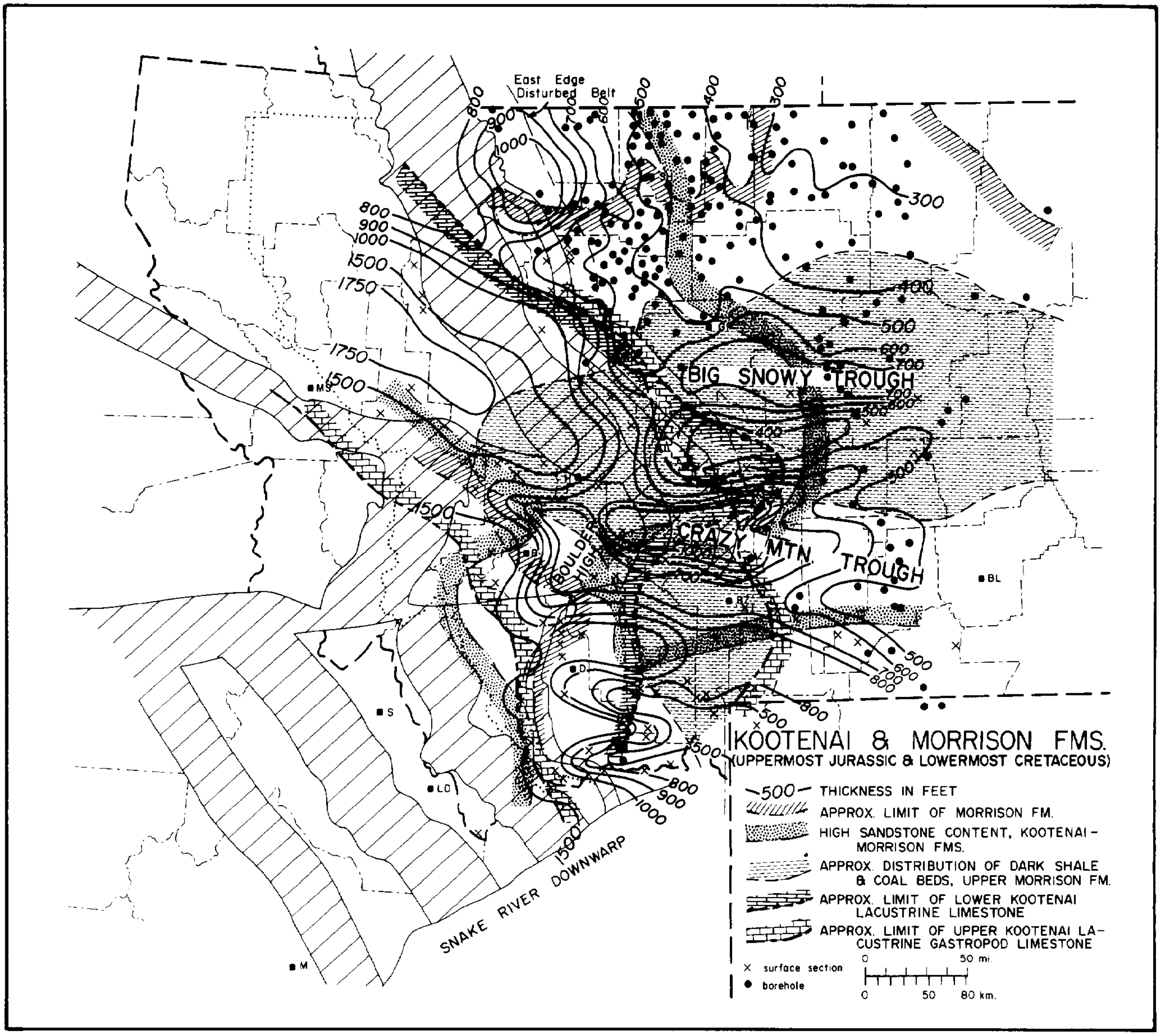

Figure 16—Thickness and general lithofacies of the Upper Jurassic Morrison and Lower Cretaceous Kootenai Formations, partly restored. Data are palinspastically restored in thrust belt.

Triassic

Rocks of Triassic age occur in a northward-thinning wedge that has a truncation offlap relationship with the underlying Permian as a result of pre-Middle Jurassic erosion (Figures 4, 7, and 14). The paleogeographic framework and source terrane characteristics of the Permian continued relatively unchanged into the Triassic, and deposition continued without evidence of marked physical interruption. Diminished influx of clastic material is indicated by the presence of finer grained clastic sediments in the Triassic beds than in the underlying Permian and Pennsylvanian facies. Triassic sediments, however, are in marked contrast to those of the underlying Permian, primarily with respect to the cessation of phosphatic, chert, and skeletal carbonate deposition that characterized the upper Paleozoic depositional facies. A primary reason for this is the apparent rapid decline of marine rock-building faunas at the close of the Permian, a phenomenon that is particularly evident in the Permian–Triassic beds of southeastern Idaho and western Wyoming (Yochelson, 1968; Peterson, 1980, 1984a, b).

Previous publications on Triassic stratigraphy and facies in and near western Montana include those of Moritz (1951) and Kummel (1954).

Jurassic: Ellis Group

Ellis Group rocks are present in the Williston basin and in almost all of the northern Rocky Mountains, as well as the entire Rocky Mountain shelf area. These beds were deposited in three well-documented transgressive-regressive cycles of Middle and Late Jurassic age (Sawtooth–Piper, Rierdon, and Swift cycles) that are a product of worldwide

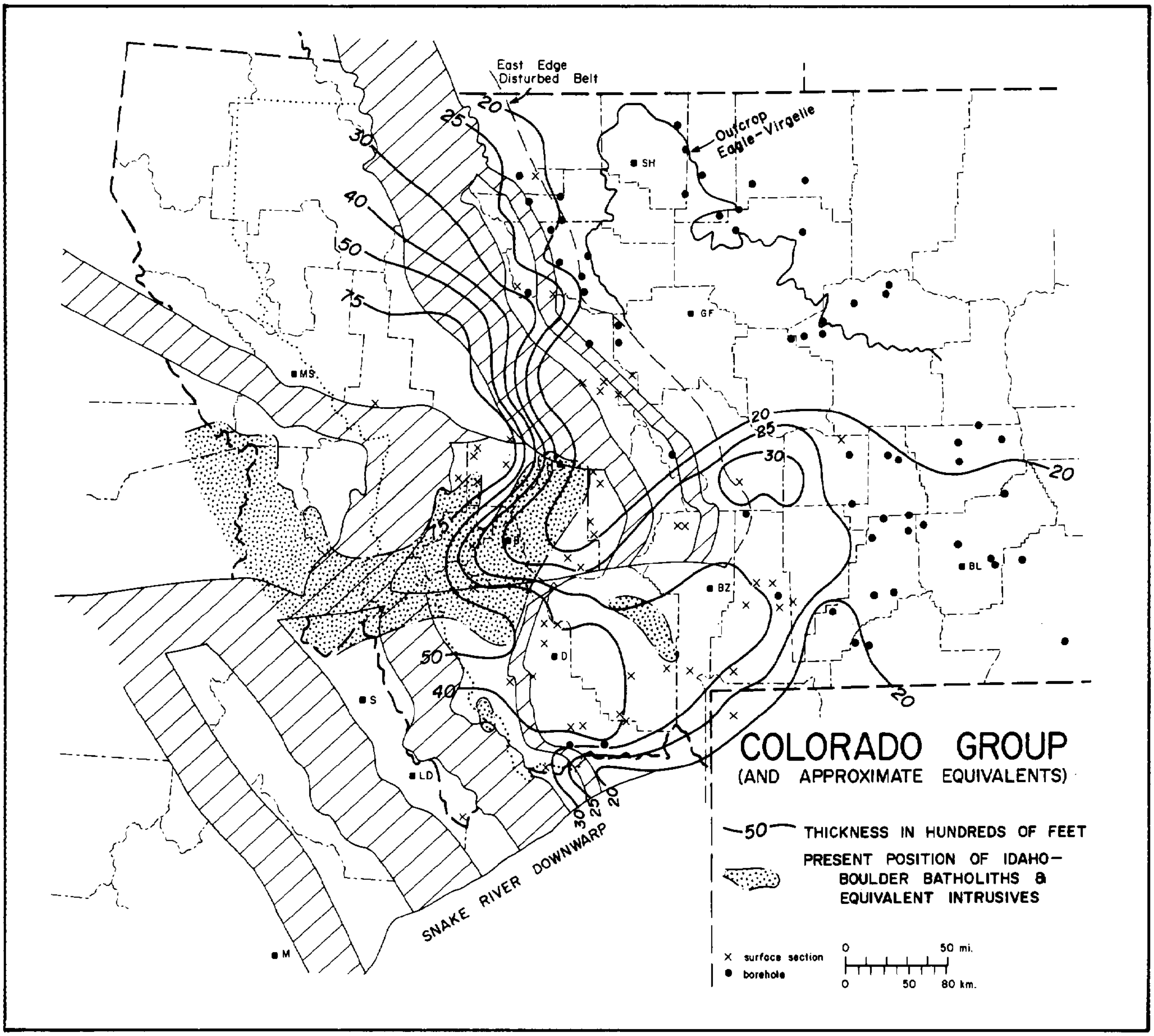

Figure 17—Thickness of Cretaceous Colorado Group and equivalent rocks, partly restored. Data are palinspastically restored in thrust belt. Present-day position (unrestored palinspastically) of Idaho and Boulder batholiths is shown.

transgressive events, which in the northern hemisphere represent incursion of northern (boreal) seas onto the continental shelf areas of that time. Rocks of Late Triassic and Early Jurassic age are not present in the northern Rocky Mountains. In Montana, Middle Jurassic basal transgressive units truncate underlying rocks ranging in age from Mississippian to Early Triassic.

Ellis paleogeography in the northern Rocky Mountains (Figure 15) was apparently dominated by the Belt Island complex of gentle uplifts. Evidence for this comes from various parts of the Ellis Group being thinned or absent because of erosion or nondeposition; this occurred during all or part of the three main depositional cycles. Vestiges of the Paleozoic and early Mesozoic paleostructural features were apparently present during Jurassic time but were much subdued. These include the Big Snowy trough and possibly the Crazy Mountain and Snowcrest troughs, all of which show minor thickening of Ellis deposits (Figure 15). Rejuvenated growth of the Sweetgrass arch is reflected by the thinning and absence of some Ellis units coinciding with the trend of the arch. A major clastic source area appeared in southwestern Montana and adjacent Idaho at this time approximately in the position of the Paleozoic Lemhi arch (Figures 2 and 15). This event probably reflected the beginning of tectonic activity that ultimately resulted in development of the western North American thrust and fold belt. The Boulder high, which at this time may have represented initiation of the igneous and tectonic event that culminated with emplacement of the Boulder batholith, retained its expression as part of an eastward projection off the Lemhi arch and formed a part of the Belt Island complex.

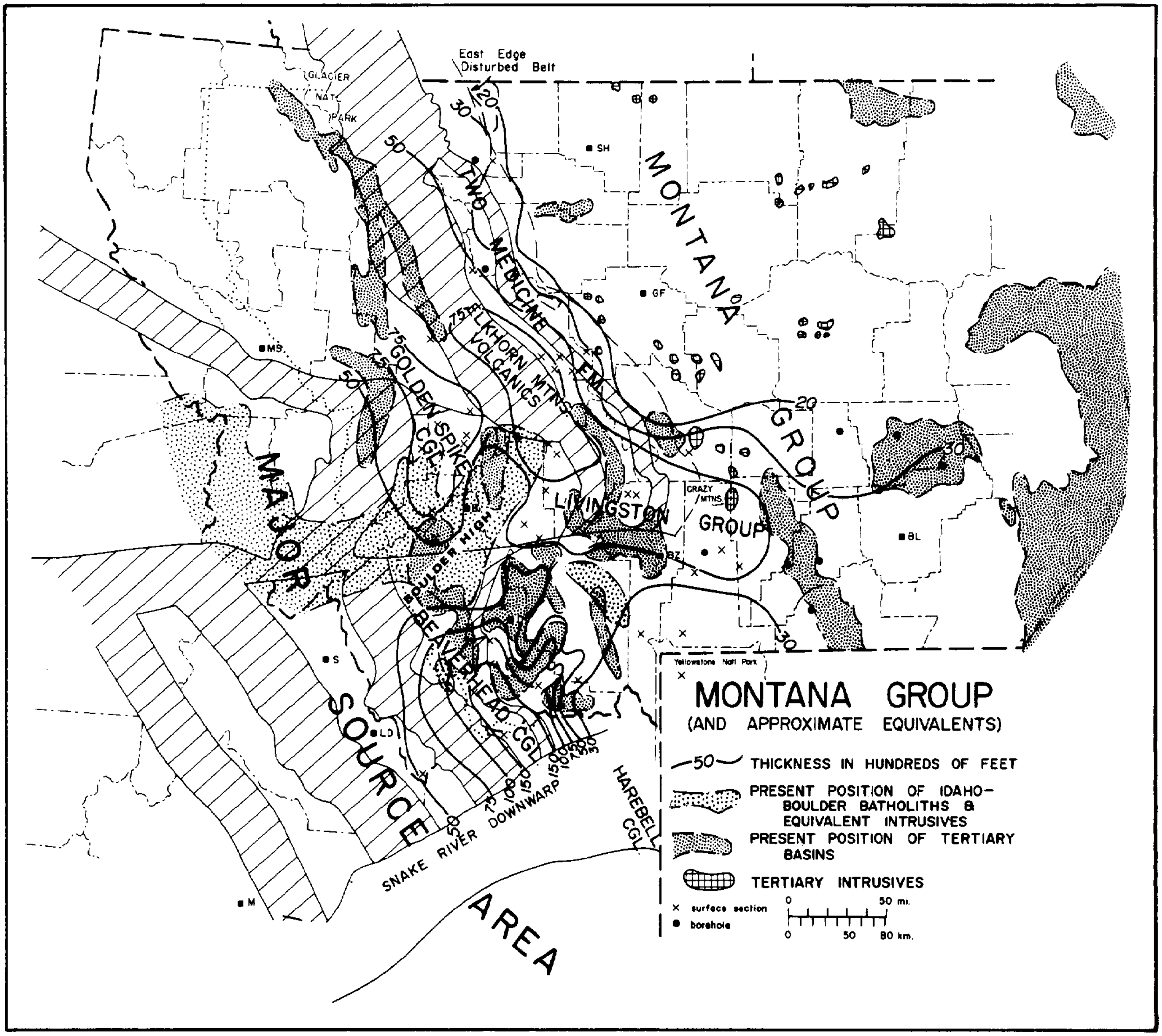

Figure 18—Thickness of Cretaceous Montana Group and equivalent rocks, partly restored. Data are palinspastically restored in thrust belt. Present-day positions (unrestored palinspastically) of Idaho and Boulder batholiths and Tertiary basins are shown.

Marine Jurassic rocks are dominated by fine- to medium-grained clastic sediments that originated mainly from the western source area. Significant carbonate deposits are present in the lower part of the Upper Jurassic sequence (Rierdon Formation), which are a northern extension of the thick Twin Creek Limestone facies of southeastern Idaho. These beds represent the final deposition of important marine carbonate sediments on the Rocky Mountain shelf. Marine clastic deposits of the Swift Formation grade into the basal beds of the overlying nonmarine Morrison Formation without evidence of unconformity.

Detailed analyses of the marine Jurassic sequences of Montana include those by Cobban (1945), Imlay (1945, 1957, 1980), Imlay et al. (1948), Moritz (1951, 1960), Schmitt (1953), Nordquist (1955), and Peterson (1957, 1972a, b).

Upper Jurassic–Lower Cretaceous: Kootenai and Morrison Formations

The Upper Jurassic Morrison Formation, the Lower Cretaceous Kootenai Formation and their equivalents are 500 ft (150 m) to more than 1500 ft (450 m) thick in western Montana (Figure 16). These beds represent an interruption of marine deposition in the western interior United States that affected the entire Rocky Mountain shelf. During this time, the shelf was covered with a blanket of varicolored clays, silts, sands, and continental lacustrine deposits. Much of the fine-grained fraction of this sedimentary blanket may have originated from fallout of fine-grained volcanic debris generated in the western North American volcanic field where important batholith emplacement occurred at this time (Stokes, 1950; Moberly, 1960; Peterson, 1966, 1972a, b; Suttner, 1969; Suttner et al., 1981). Prominent fluvial channel sand and gravel deposits are present in several parts of the Morrison and Kootenai section and their equivalents in all areas of the Rocky Mountains, and a widespread blanket deposit of boulder, gravel, and coarse sand material is commonly present at the base of the Kootenai and its equivalents on the Rocky Mountain shelf. A western source for this material is well documented, and its depositional environment has been the subject of much discussion (Stokes, 1950; Moberly, 1960; Peterson, 1966, 1972a, b; Suttner, 1969; Walker, 1974; Suttner et al., 1981).

Lacustrine limestone beds are frequently found interbedded with Morrison clastic beds, and evidence has been presented for relatively widespread, probable lacustrine dark shale and coaly deposits in the uppermost Morrison Formation, which generally cover the area of the central Montana trough (Peterson, 1966, 1972a, b; Walker, 1974). Relatively thick, widespread lacustrine limestone beds are also present in the lower and upper parts of the Kootenai Formation and equivalent beds in western Montana, southeastern Idaho, western Wyoming, and central Utah (Holm, et al., 1977). These deposits probably represent widespread north–south-oriented lake systems that formed in the foredeep area of the western tectonic belt that was progressively increasing in activity by Late Jurassic–Early Cretaceous time. At times of lower clastic influx or higher precipitation, the subsiding foredeep filled with nonmarine waters that persisted long enough to allow deposition of thick limestones containing ostracods, charaphytes, and gastropods. The upper part of the Kootenai grades upward from the lacustrine gastropod limestone beds and equivalent clastics into the lower part of the Blackleaf Formation, which contains the continental and nearshore marine sediments of the initial Cretaceous marine transgression.

Detailed studies of Morrison and Kootenai stratigraphy and facies distribution in western Montana include those by Peck (1941, 1956), Yen (1949, 1951), Moberly (1960), Suttner (1968, 1969), Paine (1970), Walker (1974), James and Oaks (1977), Holm et al. (1977, and Suttner et al. (1981).

Cretaceous: Colorado Group and Equivalent Rocks

Post-Kootenai Cretaceous rocks were deposited in a sequence of at least five transgressive–regressive cycles: the Dakota–Mowry and Frontier-Niobrara cycles, which make up the Colorado Group, and the Eagle–Claggett, Judith River–Bearpaw, and Fox Hills–Lance–Hell Creek cycles, which make up the Montana Group. In western Montana, all of these marine cycles grade westward into coarse-grained nonmarine facies that progressively increase in coarseness and content of volcanic material in the younger Cretaceous units (Figures 4, 5, 6, 7, and 17). Major tectonic elements affecting these sedimentary distribution patterns were the progressive growth and eastward spreading of the western Montana thrust and fold belt and the emplacement of the Upper Cretaceous Idaho and Boulder batholith systems and their satellite intrusives.

The Colorado Group comprises a sequence of well-defined, fossiliferous, marine transgressive–regressive cyclic clastic deposits (Dakota through Niobrara formations). These units grade westward into a complex, intertonguing marine and nonmarine facies and finally a nonmarine, clastic, nonfossiliferous, partly volcanic sequence. The main belt of marine to nonmarine transition occurs roughly in the vicinity of the central and northern parts of the western Montana disturbed belt where the eastern open marine section grades westward into the Blackleaf and Marias River formations and their equivalents (Figures 5 and 6). Volcanic debris is common in both these units, which as pointed out by McMannis (1965), may be related to the early stages of emplacement of the Idaho batholith and associated volcanic activity. In southwestern Montana, the marine units of the Colorado Group extend somewhat farther west than in the west-central and northern areas.

Thickness of the Colorado Group and equivalents increases westward from less than 2000 ft (600 m) east of the overthrust belt in north-central Montana to more than 7500 ft (2300 m) in west-central Montana (Figure 17). Part of the section is removed by Cenozoic erosion along a broad belt roughly corresponding with the trend of the Sweetgrass arch and Little Belt Mountains. This thinning trend extends southwestward for some distance across the approximate position of the Boulder high, indicating continued growth of this paleostructural feature. The section is entirely removed by erosion or nondeposition in northwestern Montana, which in large part became a major clastic source area during Cretaceous time. A general thickening trend extends eastward from the central overthrust belt approximately in the position of the central Montana trough. Continued tectonic activity in westernmost Montana and adjacent Idaho expanded the size and elevation of the western source terrane and thus increased the influx of coarse clastic material by high gradient streams off the western highland. Some foredeep lacustrine deposits are also reported in part of the Colorado Group section in west-central Montana. Volcanic debris is common in parts of the sequence.

Detailed studies of the Colorado Group include those by Cobban (1951), Gwinn (1965), Cannon (1966), Schwartz (1972), Mudge (1972), McGookey et al. (1972), and Suttner et al. (1981)

Cretaceous: Montana Group and Equivalent Rocks

Five major facies of thick, coarse-grained nonmarine facies dominate Montana Group stratigraphy in western Montana (Figures 4, 5, 6, 7, and 18).

The *Beaverhead Conglomerate facies* is a complexly intertonguing facies, more than 15,000 ft (4500 m) thick, of quartzite and limestone-boulder conglomerate and sandstone, which according to Ryder and Scholten (1973) probably ranges in age from middle Colorado to Paleocene or Eocene. The source of some of the material was the rapidly expanding highland to the west associated with the evolving overthrust belt (Figures 4 and 18), but local sources were also present east of the overthrust belt.

The *Golden Spike facies* is at least 7500 ft (2300 m) thick and is similar in nature to the Beaverhead Conglomerate facies. Part of the Golden Spike could originally have been a northern continuation of the Beaverhead Conglomerate facies, and most of it may have been removed by subsequent erosion (Figures 5 and 18) (Gwinn and Mutch, 1965). A substantial portion of the Golden Spike facies, however, was probably derived from the Boulder high (Figure 18).

The *Elkhorn Mountains volcanics* occupy the frontal position of the overthrust belt in the vicinity of Helena and Butte and intertongue westward with part of the Golden Spike facies (Figures 5 and 18). This facies is a complex pile of volcanic flows, tuffs, breccias, and tuffaceous sandstone and mudstone that intertongues with and grades eastward into units of the Livingston Group (McMannis, 1955, 1965; Roberts, 1965; Skipp and McGrew, 1977). The origin of the material is closely related to the emplacement of the Boulder batholith complex (Smedes, 1966; Robinson et al., 1969).

The *Livingston Group* is more than 5000 ft (1500 m) thick and occupies the Crazy Mountain basin and adjacent area. The Livingston is a complex of mostly nonmarine volcaniclastic sandstone and mudstone containing a few volcanic flows and tuffs. The lower part intertongues rapidly eastward with marine units of the Montana Group on the east flank of the Crazy Mountain basin (Figures 5 and 18). The upper two-thirds of the Livingston intertongues with nonmarine beds of the Hell Creek and Lance formations (Roberts, 1965).

The *Two Medicine Formation* is preserved in outcrop remnants, some of which are at least 5000 ft (1500 m) thick, in the frontal zone of the northern disturbed belt near and south of Glacier National Park. Eastward in the subsurface, the nonmarine Two Medicine thins to 2000 ft (600 m) or less and grades into marine units of the Montana Group along the west flank of the Sweetgrass arch (Schwartz, 1972). To the south it intertongues with the Elkhorn Mountains volcanic facies (Figures 6 and 7).

The overall pattern of these facies relationships indicates that during Montana time, coarse boulder conglomerate debris that was derived primarily from erosion of Precambrian and Paleozoic rocks was transported eastward from the rising highland belt associated with the rapidly expanding western thrust and fold belt. Volcanic debris and flows were concentrated around the periphery of the Boulder high, which during Cretaceous (Montanan) time was the main site of emplacement of the Boulder batholith and its associated intrusive rocks. The marine Cretaceous seaway apparently transgressed westward into this complex of volcanic and conglomeratic debris with frequent rapid fluctuations of the western shoreline. At the same time, much volcanic ash and tuff fallout was incorporated with normal shallow water marine deposits to the east, which resulted in the common layers of bentonite and bentonitic shale characteristic of the Colorado and Montana Group beds in central and eastern Montana.

Cretaceous sedimentation in western Montana closed with the deposition of the nonmarine St. Mary's River Formation in the Glacier Park area and the Adel Mountains volcanic rocks to the south. In southwestern and west-central Montana, rocks of this age have been almost entirely removed by subsequent erosion, but age equivalents are probably preserved in the upper part of the Beaverhead Conglomerate and the Sphinx Conglomerate near Yellowstone National Park.

Detailed studies of Montana Group facies and correlative rocks in western Montana include those by McLaughlin and Johnson (1955), Klepper et al. (1957), Weimer (1960, 1961), Roberts (1963, 1965), Gwinn (1965), Gwinn and Mutch (1965), Viele and Harris (1965), Simms (1967), Wilson (1967), Ryder and Ames (1970), Ryder and Scholten (1973), McGookey et al. (1972), and Skipp and McGrew (1977). Studies of the age and emplacement processes of the Boulder batholith complex include those by Weeks and Klepper (1954), Klepper et al. (1957), Chapman et al. (1955), Smedes (1966), Burfiend (1967), Robinson et al. (1969), Hamilton and Myers (1974), and Hyndman et al. (1975).

Tertiary

Sedimentation patterns of early Tertiary time reflect the continuation without significant interruption of the closing Cretaceous paleogeographic framework and depositional basins of the northern Rocky Mountains. The major change influencing depositional environments was the final withdrawal of Upper Cretaceous marine waters from the Rocky Mountain shelf. Marine deposition was replaced by widespread lacustrine and fluvial deposition across the emergent, gently sloping Late Cretaceous sea bottom, and extensive lacustrine and coastal swamp deposition of coaly beds and clastics occurred in the more active basinal areas of the shelf. These deposits make up the Hell Creek Formation and overlying Fort Union Group of the Montana plains (Figures 5 and 6). In western Montana, remnants of equivalent age rocks are scarce, but they are probably represented in the upper part of the Beaverhead and Sphinx conglomerates of southwestern Montana and the nonmarine Willow Creek Formation and Adel Mountains volcanics south of Glacier National Park.

By late Eocene time, the downfaulted continental Tertiary basins had been relatively well developed, and the remainder of Tertiary sedimentation was involved with basin-fill deposition in the many downwarped valley areas of present-day western Montana (Figure 18). Middle to Late Tertiary fluvial and lacustrine beds reach thicknesses of as much as 5000–10,000 ft (1500–3000 m) in several of the basins (C. A. Balster, personal communication, 1985). Burfiend (1967) reported gravity data that indicate that some of the valleys between Bozeman and Butte are filled with at least 5000 ft (1500 m) of unconsolidated material. According to Kuenzi and Fields (1971), the valley fill in the Jefferson basin east of Butte contains an upper Eocene or Oligocene to lower Miocene, fine-grained clastic sequence overlain by a Miocene and Pliocene sequence of coarser material that are separated by a marked unconformity. Similar stratigraphic sequences are reported in other Tertiary valleys of western Montana, and evidence suggests that the unconformity may be a regional feature common to valley fill history in many parts of western Montana (Robinson, 1961; Kuenzi and Richard, 1969). Many of the late Tertiary basins probably were controlled by rejuvenation of Precambrian tectonic features in western Montana (O'Neill and Lopez, 1985).

The age and mechanisms of thrusting and other tectonic activity that shaped the final form of western Montana are discussed in detail by many authors, including Alpha (1955, 1958), Poulter (1956, 1958), Scholten (1956, 1960, 1967, 1968), Reid (1957), Foose et al. (1961), Mutch (1961), Gwinn (1961), Childers (1963), Armstrong and Oriel (1965), Weidman (1965), McGill (1965), Smith and Barnes (1966), Scholten and Ramspott (1968), Mudge (1970), Klepper et al. (1971), Bregman (1971), Tysdal (1976), Skipp and Hait (1977), Ruppel (1978, 1982,), Ruppel and Lopez (1984), Hadley (1980), and Woodward (1980).

PETROLEUM GEOLOGY

Many of the components for petroleum generation, accumulation, and preservation are present in western Montana. Much of the geologic section has a shallow marine origin and contains variable lithologic facies that include substantial thicknesses of potential reservoir rock, source rock, and cap rock, which in many cases are closely associated under favorable stratigraphic trapping conditions. The presence of paleostructural elements and shelf marginal belts, which persisted throughout most of the depositional history here, provided a combination of favorable conditions for the localized development of good reservoir rock facies and the continuous growth of a stratigraphic and structural framework favorable to early migration and trapping of hydrocarbons at the time of generation. Factors that may negatively affect hydrocarbon accumulation include deep burial and tectonic alteration of clastic reservoir rocks; excessive thermal effects on source rocks; and destruction of early traps by stratigraphic and structural events in parts of the area, such as excessive burial depths and the complex structural and igneous activity during the Mesozoic and Tertiary. However, fracturing of potential reservoir rocks, particularly dolomites, by intense structural activity should be a positive factor.

Cambrian

The thick, porous Hasmark dolomite facies is of interest because of its potential for trapping petroleum, possibly nonindigenous, under suitable structural conditions. Evidence for adequate Cambrian source rocks is scarce, and burial depths beyond the postmature stage of organic diagenesis have affected these rocks in much of western Montana (Figures 4, 5, 6, and 7).

Devonian

Jefferson Formation dolomite beds are porous almost everywhere and commonly show indications of petroleum in the form of staining or odor. This thick reservoir facies is of major interest, although important petroleum accumulations have not yet been found in these beds in central to western Montana. Potentially good source rock beds are present in the Devonian–Mississippian Milligen Formation of east-central Idaho and southwestern Montana and in the thinner, partially equivalent, Bakken–Sappington facies in western Montana. Remnants of these facies may be present beneath thrusts in parts of central western or northwestern Montana. According to Perry et al. (1981), these beds have been thermally altered to postmaturity west of the Medicine Lodge thrust. In much of western Montana, however, these rocks probably have not been buried to depths beyond the mature stage (Figures 4, 5, 6, and 7).

Mississippian

Some porous dolomite units are present within the Madison bank carbonate facies in west-central to southwestern Montana. These beds show indications of petroleum at some outcrop localities. The Sun River Dolomite facies, the major Paleozoic reservoir in the Sweetgrass arch region, covers a broad area of northwestern and west-central Montana (Figure 11) and is of major interest for future exploration. The Bakken–Sappington source rock facies is about 1000 ft (300 m) or more stratigraphically below the Sun River, but along with dark shale and argillaceous limestone beds in the lower Lodgepole and equivalents, it is potentially a good source rock facies in much of the area. Except for parts of southwestern and west-central Montana, these beds probably have not been buried to excessive depths, and in shallow burial areas they may never have been heated beyond the early stages of organic maturity (Figures 4, 5, 6, and 7). The highly organic-rich shale beds of the Heath Formation extend westward and southwestward from the Big Snowy trough and are present in a large part of the central and southwestern thrust belt and in east-central Idaho (Figure 12). This source rock facies has been deeply buried in east-central Idaho and in the Snowcrest trough, but in the remainder of western Montana burial depths probably did not reach the postmature stage.

Pennsylvanian–Permian

The Quadrant Formation and its equivalent shelf sandstone sequence are a major reservoir facies in much of the northern Rocky Mountains, particularly in Wyoming. These beds have been deeply buried and are quartzite or dolomite in most of southwestern and west-central Montana, but interbedded porous sandstones of probable eolian origin are present in the central and southwestern thrust belt and adjacent area (D. Saperstone, personal communication, 1984). The overlying Permian sandstone and carbonate beds generally are not very porous and are of only moderate interest as reservoirs. The highly organic-rich Retort and Meade Peak phosphatic shale beds have excellent source rock qualities in much of the northern Rocky Mountain region, including southwestern Montana. In some places these rocks are not thermally mature, such as in the Tendoy thrust area (Claypool et al., 1978; Perry et al., 1981, 1983), but in the Snowcrest trough and in some of the Tertiary basins, or under thrust sheets, they may have reached thermal maturity by late Mesozoic time (Figures 4 and 5).

Triassic–Jurassic

Reservoir and source rock quality of Triassic beds and of marine beds of the Jurassic appear to be low in west-central to southwestern Montana. However, in the thrust belt west of Great Falls, the Sawtooth Formation includes dark marine shale beds that are phosphatic in places; these are of interest as source rocks and may be the source for some of the Sweetgrass arch petroleum. Dark marine shale beds are a minor component of the Rierdon and Swift formations in the same area, and these, along with dark shale and coaly lacustrine beds in the upper Morrison Formation, deserve some attention for moderate to low potential.

Cretaceous

The intertonguing belt of marine and continental facies of the Colorado Group falls generally within and just east of the overthrust belt. Good clastic nearshore continental and marine reservoir sandstone beds and dark continental or marine shales are present within much of this facies. Within and adjacent to the thrust belt, these beds may have been buried deeply enough to reach thermal maturity (Figures 4, 5, 6, and 7). The similar continental to marine transition facies of the Montana Group generally occurs farther east. In most places, this sequence probably has not been buried to thermal maturity depths, but potential for biogenic gas generation may have been good if trapping conditions were favorable in these beds and in beds of the Colorado Group (Rice and Shurr, 1978; Clayton et al., 1982).

Tertiary

Discontinuous alluvial sandstone bodies, some with good porosity, are common in the early to middle Tertiary beds of the western Montana Tertiary basins, and some carbonaceous to coaly beds are also present. Except for the deeper parts of these basins, most of this section has not been buried to thermal maturity, but the possibility of small to moderate sized biogenic gas accumulations, mostly in stratigraphic traps, deserves consideration in future exploration programs.

Summary of Petroleum Geology

The western Montana overthrust belt contains a less complete and generally thinner stratigraphic section and less extensive potential source rock facies than that in the overthrust belt of western Wyoming, southeastern Idaho, and northern Utah where major petroleum accumulations have been discovered in recent years. The western Montana area is also characterized by extensive Mesozoic and Cenozoic igneous activity, in contrast with western Wyoming and southeastern Idaho. In addition, the high degree of structural complexity and surface exposure of much of the potentially favorable rock section and the detrimental effects of deep burial of Paleozoic rocks all detract from the petroleum potential of the region. The data presented by Claypool et al. (1978) and Perry et al. (1981, 1983), however, indicate that the Phosphoria and the highly organic-rich Big Snowy source rock facies have not reached the full maturity thermal stage in the Tendoy thrust region. Thus, windows of less rigid thermal effects may be preserved in places, particularly in the vicinity of persistent paleostructural highs. The palinspastically restored stratigraphic cross section in southwestern Montana (Figure 4) indicates that the middle and upper Paleozoic section had not been buried beyond the early mature stage by the end of Jurassic time. This was probably near the time of maximum burial depth of Paleozoic beds in this area. During Late Jurassic and later deposition, much of this area underwent relatively continuous uplifting as the Mesozoic fold and thrust system evolved. During this time, the Paleozoic section was relatively continuously uplifted and finally partly eroded, and the thermal gradient burial cycle may have either remained dormant or reversed in direction, resulting in the preservation of thermally less mature windows.

A large area of unexplored terrain remains in western Montana. Burial depths are within a favorable range in significant parts of the area, and a large volume of favorable reservoir facies is present. At this time, geologic appraisal and the drilling history of the area continues to point primarily toward its gas potential, with low probability for significant oil resources.

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