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Chapter CC

Pennsylvanian and Early Permian Paleogeography of the Uinta-Piceance Basin Region, Northwestern Colorado and Northeastern Utah

By SAMUEL Y. JOHNSON, MARJORIE A. CHAN, and EDITH A. KONOPKA

A multidisciplinary approach to research studies of sedimentary rocks and their constituents and the evolution of sedimentary basins, both ancient and modern

U.S. GEOLOGICAL SURVEY BULLETIN 1787

EVOLUTION OF SEDIMENTARY BASINS—UINTA AND PICEANCE BASINS

U.S. DEPARTMENT OF THE INTERIOR
MANUEL LUJAN, JR., Secretary

U.S. GEOLOGICAL SURVEY
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Pennsylvanian and Early Permian Paleogeography of the Uinta-Piceance Basin Region, Northwestern Colorado and Northeastern Utah

By Samuel Y. Johnson[1], Marjorie A. Chan[2], *and* Edith A. Konopka[3]

Abstract

The Uinta-Piceance basin region of northwestern Colorado and northeastern Utah includes parts of four major sedimentary provinces that were active during the late Paleozoic ancestral Rocky Mountain orogeny: the Eagle basin, the northern part of the Paradox basin, the southern Wyoming shelf, and the southeastern part of the Oquirrh basin. Depositional patterns in these sedimentary provinces were controlled to varying degrees by eustatic and climatic fluctuations (forced by expansion and contraction of late Paleozoic continental ice sheets), tectonism, and sediment supply. Four sets of paleogeographic maps illustrate the major changes in paleogeography and depositional patterns associated with repetitive transgressions and regressions. In general, clastic deposition (mostly sandstone in fluvial, deltaic, and eolian systems) dominated during regressions, whereas deposition of marine limestone and clastic rocks characterized transgressions. The most arid climates are correlated with regressions.

Morrowan and lower Atokan strata throughout most of the Uinta-Piceance basin region consist of fine-grained clastic rocks (regressive deposits) and more abundant limestone (transgressive deposits). Significant Morrowan and early Atokan tectonic uplifts include the Front Range and Sawatch uplifts.

The late Atokan to Desmoinesian history of the region is characterized by unroofing of the Front Range and Sawatch uplifts and rapid rise of the Uncompahgre uplift. In the central parts of the Eagle and Paradox basins, tectonic activity resulted in restricted circulation and low clastic sediment supply, creating conditions suitable for evaporite deposition during regressions. Regressive deposition was also characterized by significant progradation of eolian sands across the Wyoming shelf en route to the Oquirrh basin. Limestone deposition was dominant throughout the region during transgressions.

During the Missourian and Virgilian, progradation of clastic sediments into the central parts of the Eagle and northern Paradox basins resulted in termination of evaporite deposition. During regressions, fluvial and eolian deposition dominated in the Eagle basin, and sabkha and (or) shallow marine deposition dominated in the northern Paradox basin. The Wyoming shelf continued to serve as a conveyor belt of clastic sediment to the Oquirrh basin, which was characterized by a transition from shallow- to deep-water deposition. Deposition of limestone during transgressions was limited to the western part of the Uinta-Piceance basin region and a small area in the eastern Eagle basin.

In the early Wolfcampian, the Emery uplift became fully or mostly submerged, and the Paradox basin ceased to be a discrete geomorphic element. Deep-water clastic deposition in the Oquirrh basin continued from the latter part of the Missourian-Virgilian. Missourian-Virgilian depositional patterns in the Eagle basin and on the Wyoming shelf also continued into the early Wolfcampian. Deposition of limestone during transgressions was limited to the southwestern part of the Uinta-Piceance region.

Regionally, rates and magnitude of subsidence were greatest in the Oquirrh basin, intermediate and very similar in the Eagle and northern Paradox basins, and lowest on the Wyoming shelf. In all four sedimentary provinces, rates of subsidence were lowest in the Early Pennsylvanian, highest in the Middle Pennsylvanian, and intermediate in the Late Pennsylvanian and Early Permian. The timing, magnitude, and geometry of subsidence in the Uinta-Piceance basin region strongly

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[1]U.S. Geological Survey, MS 939, Box 25046, Denver Federal Center, Denver, Colorado 80225.

[2]Department of Geology and Geophysics, University of Utah, Salt Lake City, Utah 84112.

[3]Department of Geology and Geophysics, University of Wisconsin, Madison, Wisconsin 53706. Present address: 541 Fairmount Avenue, Westfield, New Jersey 0790–1358.

suggest that regional intraplate deformation reflects the overlapping influences of interactions along a more distant, convergent continental margin to the south and a more proximal, transform-fault (?) continental margin to the west.

INTRODUCTION

The rectangular Uinta-Piceance basin region of northwestern Colorado and northeastern Utah encompasses the Laramide (latest Cretaceous to Paleogene) Uinta and Piceance basins and surrounding uplifts (fig. 1). The area is mostly underlain by Phanerozoic sedimentary rocks that represent at least five different phases of basin evolution (Johnson, Tuttle, and others, 1990). Pennsylvanian and lower Wolfcampian rocks were deposited during one of these phases, generally referred to as the ancestral Rocky Mountain orogeny. During this orogeny, the Uinta-Piceance region included parts of four major sedimentary provinces: the Eagle basin, the Paradox basin, the Wyoming shelf, and the Oquirrh basin (fig. 2). Episodic deposition also occurred on the Emery uplift and the Callville shelf. Pennsylvanian and lower Wolfcampian rocks are widespread in the subsurface of the Uinta-Piceance region, but outcrops are restricted to a few areas (fig. 1). Late Paleozoic paleogeographic reconstructions of all or significant parts of the Uinta-Piceance region include those by Hallgarth (1967), Mallory (1972, 1975), Rascoe and Baars (1972), Chronic (1979), Welsh and Bissell (1979), and DeVoto and others (1986).

Recent detailed sedimentologic studies (for example, Hite and Buckner, 1981; Driese and Dott, 1984; Johnson and others, 1988; Schenk, 1989; Karachewski, 1992)

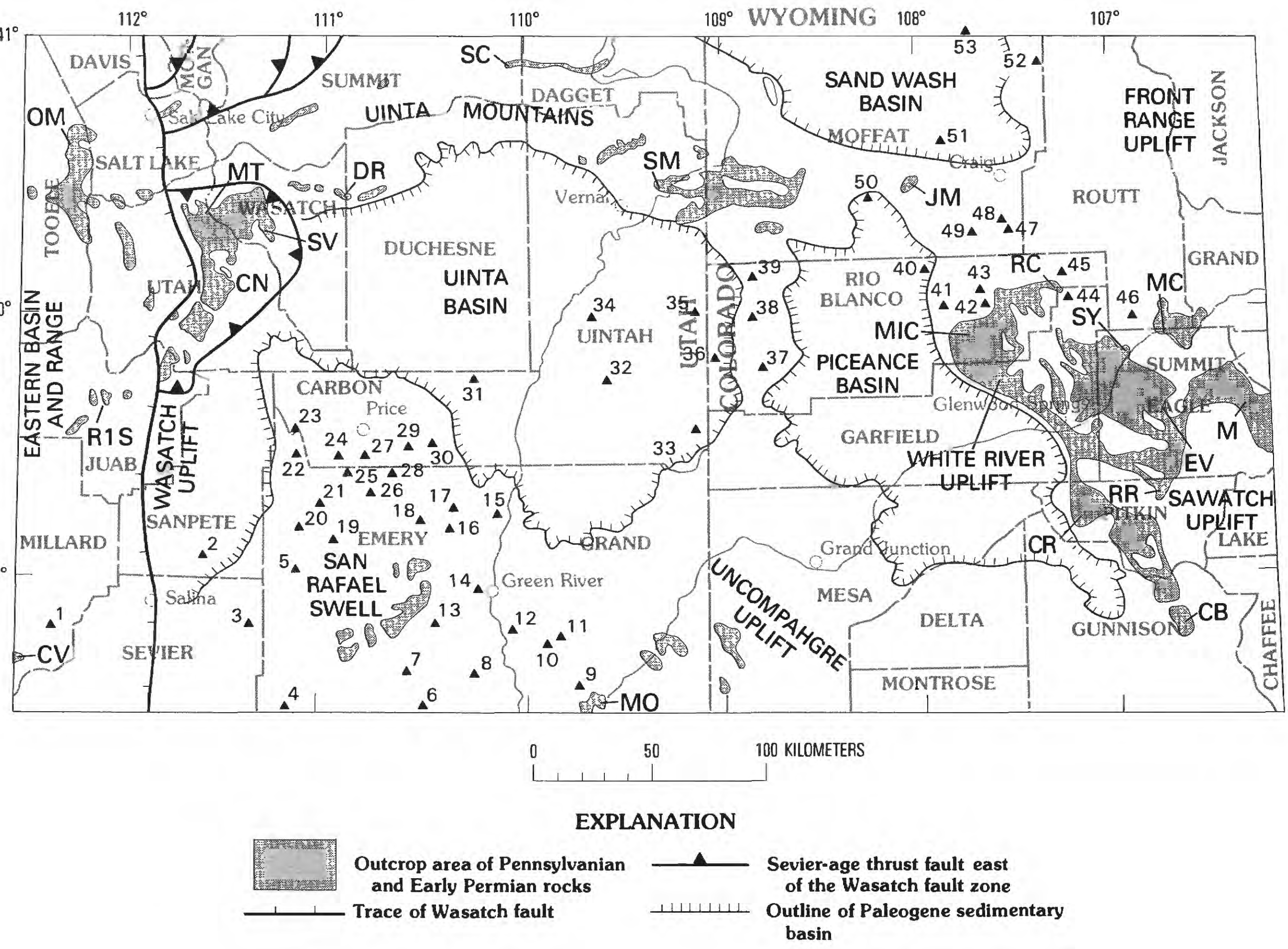

Figure 1. Map showing the Uinta-Piceance basin region in northwestern Colorado and northeastern Utah. Abbreviations for outcrop locations cited in text: CB, Crested Butte, Colorado; CN, Charleston-Nebo thrust sheet; CR, Crystal River, Colorado; CV, Cove Fort, Utah; DR, Duchesne River, Utah; EV, Eagle Valley, Colorado; JM, Juniper Mountain, Colorado; M, Minturn, Colorado; MC, McCoy area, Colorado; MIC, Miller Creek, Colorado; MO, Moab, Utah; MT, Mount Timpanagos, Utah; OM, Oquirrh Mountains, Utah; RC, Ripple Creek, Colorado; RR, Ruedi Reservoir, Colorado; R1S, Route 148 southwest, Utah; SC, Sols Canyon, Utah; SM, Split Mountain, Utah; SV, Strawberry Valley, Utah; SY, Sylvan, Colorado. Numbered triangles show locations of boreholes cited in text and listed in the appendix.

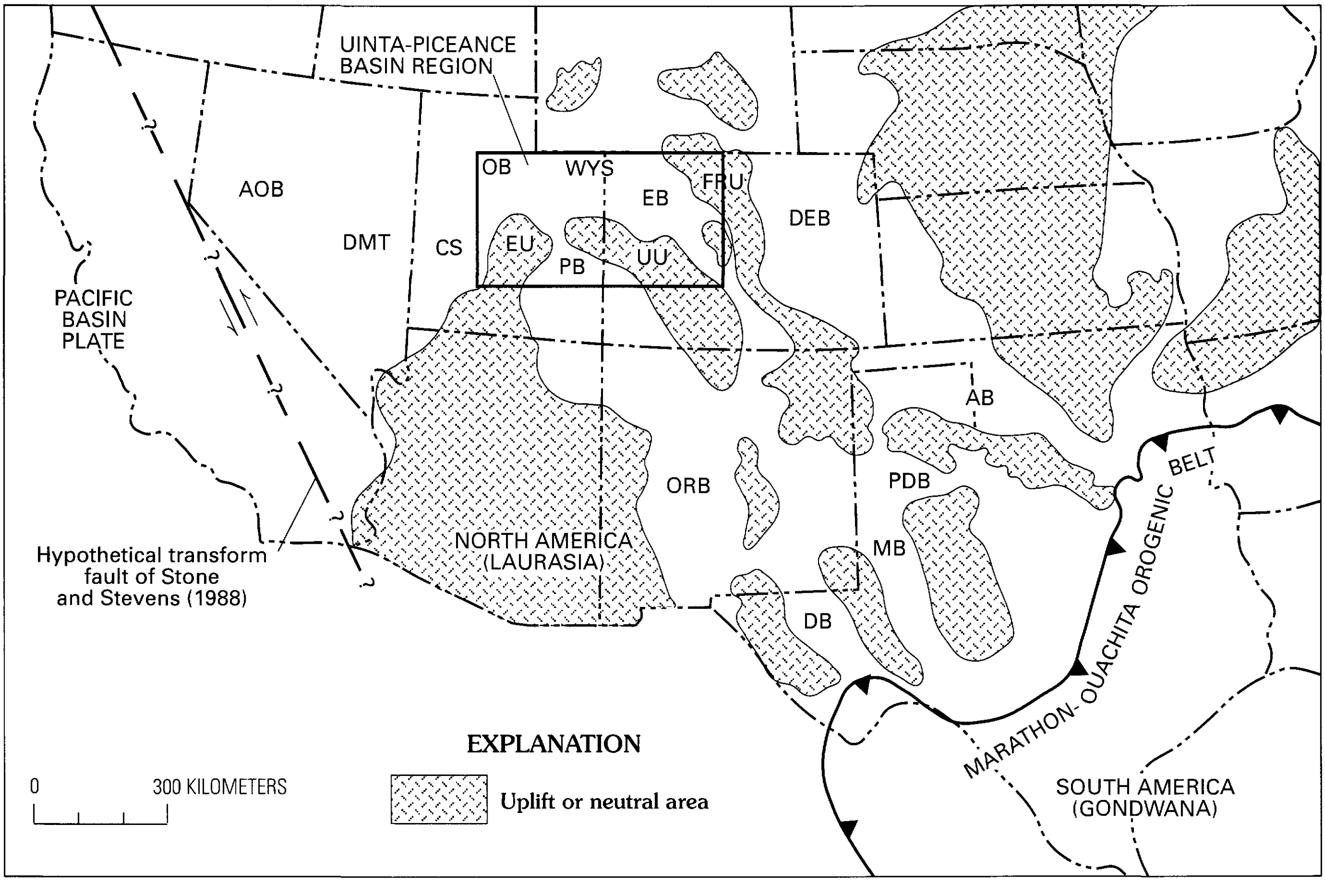

Figure 2. Schematic map showing paleogeography of the Uinta-Piceance basin region, adjacent basins and uplifts of the ancestral Rocky Mountain orogeny, and inferred continental margins to the west and southeast. Map does not restore Mesozoic shortening and Cenozoic extension in the western United States. Abbreviations: AB, Anadarko basin; AOB, Antler overlap basins; CS, Callville shelf; DB, Delaware basin; DEB, Denver basin; DMT, Dry Mountain trough; EB, Eagle basin; EU, Emery uplift; FRU, Front Range uplift; MB, Midland basin; OB, Oquirrh basin; ORB, Orogrande basin; PB, Paradox basin; PDB, Palo Duro basin; UU, Uncompahgre uplift; WYS, Wyoming shelf. Areas characterized by minimal uplift or subsidence are considered "neutral." Modified from McKee and Crosby (1975), Ross (1986), Stone and Stevens (1988), and Smith and Miller (1990).

demonstrate that Pennsylvanian and lower Wolfcampian strata of the Uinta-Piceance region are commonly cyclic, characterized by repetitive sequences of lithofacies that represent repetitive successions of depositional environments. Comparable cyclic deposition in upper Paleozoic strata has been widely recognized globally and is generally attributed to eustatic fluctuations and associated climate changes forced by expansions and contractions of Gondwana continental ice sheets (for example, Crowell, 1978; Heckel, 1986; Ross and Ross, 1987, 1988).

Existing paleogeographic maps and reconstructions for the Uinta-Piceance region greatly homogenize geologic history by showing only the dominant lithology deposited during a particular cycle or sequence and not the variations in depositional environments and lithofacies represented in the sedimentary cycles. The maps presented herein show inferred differences between eustatic transgressive and regressive deposition and thus differ significantly from earlier paleogeographic reconstructions. Paired maps for four time intervals (Morrowan and early Atokan, late Atokan and Desmoinesian, Virgilian and Missourian, early Wolfcampian) are presented. These maps represent the duration of the ancestral Rocky Mountain orogeny and also correspond to the Absaroka I sequence of Sloss (1982). The maps are speculative because many relevant rock units have not been studied in detail (particularly in the subsurface), and they should be regarded as preliminary working models. Their purpose is to illustrate the likely large-scale regional variations within a time interval and, in so doing, to demonstrate the necessity of this type of approach for reconstructing late Paleozoic paleogeography.

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LATE PALEOZOIC TECTONISM AND REGIONAL GEOLOGY

The Uinta-Piceance basin region (fig. 1) is part of the ancestral Rocky Mountains orogenic province of the west-central United States. This province is somewhat enigmatic in that it formed in an intraplate tectonic setting distant from active continental margins (fig. 2). Kluth and Coney (1981), Kluth (1986), and Budnik (1986) suggested that deformation in this province resulted from the collisional suturing of North America and South America–Africa (the Laurasia and Gondwana plates) along the Marathon-Ouachita-Appalachian orogenic belt. Kluth (1986) showed that the timing of deformation along this suture and in the ancestral Rocky Mountains was broadly correlative. Many of the major faults in the ancestral Rockies, most prominently the Wichita megashear (Budnik, 1986), probably are reactivated zones of preexisting structural weakness. Stone (1977) and Budnik (1986) emphasized sinistral strike-slip offset and deformation on major northwest-trending structures in the ancestral Rockies, whereas Stevenson and Baars (1986) postulated dextral strike-slip offset on northwest-trending faults and sinistral offset on northeast-trending faults. Kluth (1986) suggested that ancestral Rocky Mountains deformation was in some ways similar to the intraplate Cenozoic deformation of Asia associated with its collision with India; however, fault geometry and inferred kinematics between the Asia analogue and the ancestral Rocky Mountain orogeny are significantly different.

Both Kluth (1986) and Budnik (1986) considered the western continental margin of North America as passive during the late Paleozoic and thus incapable of providing a driving force for intraplate deformation. More recently, Stone and Stevens (1988) proposed that this continental margin was a northwest-trending left-lateral transform fault (fig. 2) that had a structural trend similar to that proposed for at least parts of the ancestral Rocky Mountains. Smith and Miller (1990) similarly presented evidence for significant late Paleozoic extension in Nevada that they related partly to a western driving force. If the effects of Mesozoic shortening and Cenozoic extension are removed from the southwestern United States (following Levy and Christie-Blick, 1989), then the proposed left-lateral transform fault of Stone and Stevens (1988) was about 500–600 km from the northern Paradox basin, much closer than the distance (about 1,050–1,250 km) between the Paradox basin and the Marathon-Ouachita-Appalachian orogenic belt. Using the Neogene San Andreas fault and Basin and Range extension as a very limited analogue, diffuse extensional shearing related to transform faulting can occur hundreds of kilometers from continental margins (Atwater, 1970; Wernicke and others, 1988) and thus could have affected the ancestral Rocky Mountains.

The ancestral Front Range and Uncompahgre uplifts (fig. 2) are the major late Paleozoic orogenic highlands in the Uinta-Piceance basin region. Subsiding areas include the Eagle, Paradox, and Oquirrh basins, the Wyoming shelf, the Emery uplift, and the Callville shelf. Isopach maps showing the thickness of Pennsylvanian to Wolfcampian sedimentary deposits are in Mallory (1972, 1975), Rascoe and Baars (1972), and DeVoto and others (1986). Late Paleozoic regional stratigraphy is shown in figure 3. The ancestral Front Range and Uncompahgre uplifts, cored by Precambrian crystalline rocks, form the eastern and southwestern boundaries of the Eagle basin (fig. 2), respectively. These uplifts had considerable structural and topographic relief and were bounded by narrow complex fault zones. The ancestral Sawatch uplift formed an intrabasinal block in the southern part of Eagle basin and was probably characterized by lesser positive relief during the late Paleozoic (Johnson and others, 1988; Johnson, 1989a, b). The Emery uplift of central Utah was intermittently emergent during the Pennsylvanian to Wolfcampian; at those times it formed the northwestern boundary of the Paradox basin. The Emery uplift was never deeply eroded (Mississippian strata are preserved below an unconformity with Wolfcampian rocks), however, and thus probably had minimal topographic relief.

The Gore fault zone forms the boundary between the ancestral Front Range uplift and the Eagle basin. Reactivation of this fault at least three times during the Mesozoic and Cenozoic (Tweto, 1977) has obscured its late Paleozoic history. The Gore fault is now locally a reverse fault, but DeVoto (1980) and DeVoto and others (1986) showed it as a normal fault during the late Paleozoic. The normal-fault hypothesis suggests a local extensional environment that is consistent with the presence of interbedded volcanic rocks in the Middle Pennsylvanian Minturn Formation (fig. 3) (Koschmann and Wells, 1946) within a kilometer of the Gore fault. Alternatively, the presence of a deep late Paleozoic trough adjacent to the ancestral Front Range (Johnson and others, 1988, fig. 3) could reflect crustal loading from the east and reverse motion on the Gore fault.

The unexposed Garmesa fault zone forms the boundary between the ancestral Uncompahgre uplift and the Eagle basin (fig. 2). Stone (1977) suggested that the Garmesa fault was a late Paleozoic oblique-slip fault; Heyman and others (1986, fig. 13) showed it as a reverse fault in the late Paleozoic. Numerous intrabasinal faults (with inferred normal, reverse, and oblique-slip displacement) were also active in the Eagle basin during the late Paleozoic (for example, Stone, 1977, 1986; DeVoto, 1980; Waechter and Johnson, 1985, 1986; DeVoto and others, 1986; Dodge and Bartleson, 1986; Johnson and others, 1988; Schenk, 1989; Johnson, Tuttle, and others, 1990). These structures had significant influence on patterns of deposition and subsidence. The Eagle basin includes as much as 3,200 m of Pennsylvanian and Wolfcampian strata (Johnson and others, 1988). It merges to the northwest with the southern part of the Wyoming shelf depositional province (fig. 2). The Wyoming shelf contains a comparatively thin section (about 700–800

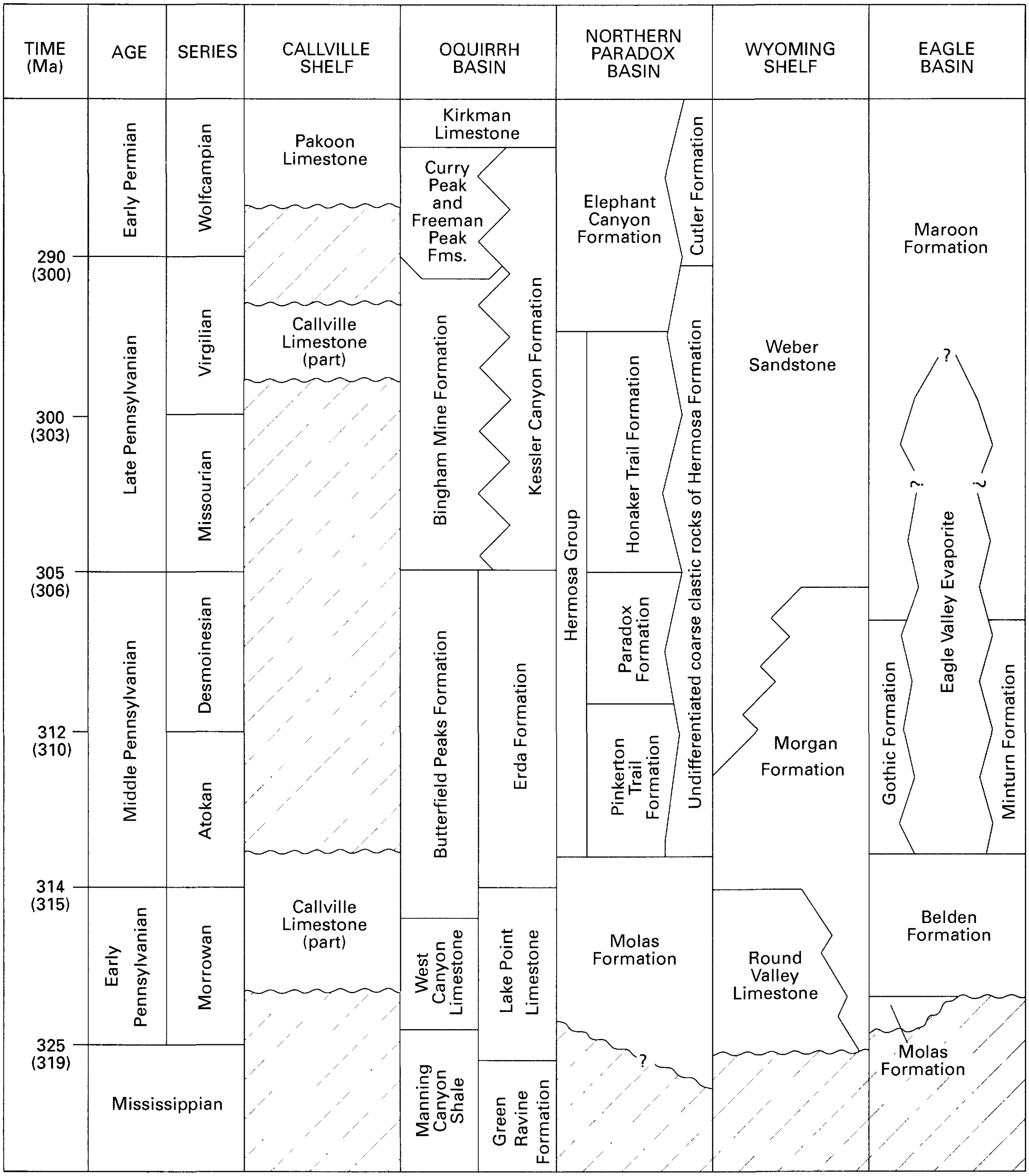

Figure 3. Chart showing Pennsylvanian and Early Permian stratigraphy of five sedimentary provinces in Uinta-Piceance region. Units may have different age ranges in areas other than those discussed here. Time scale for age and series boundaries is from Haq and Van Eysinga (1987); suggested time scale of Klein (1990) is shown in parentheses. Loope and others (1990, fig. 2) suggested a different stratigraphic nomenclature for Paradox basin in which Hermosa Group has formation status, lower part of upper member of Hermosa Formation is equivalent to Honaker Trail Formation, and upper part of upper member of Hermosa Formation and "lower Cutler beds" are equivalent to Elephant Canyon Formation. References: Callville shelf (Brill, 1963); Oquirrh basin (Tooker and Roberts, 1970; compilation of Hintze, 1988); northern Paradox basin (Baars, 1962; Mallory, 1972; Welsh and Bissell, 1979; compilation of Hintze, 1988; Baars, 1991); Wyoming shelf (Hansen, 1965; Welsh and Bissel, 1979; Driese and Dott, 1984; DeVoto and others, 1986; compilation of Hintze, 1988); Eagle basin (Mallory, 1972; DeVoto, 1980; DeVoto and others, 1986). Lower Molas Formation in the northern Paradox basin may be as old as latest Mississippian, following Merrill and Winar (1958) and suggestions of A.C. Huffman, Jr., and S.M. Condon (oral commun., 1991).

m) of Pennsylvanian to Wolfcampian strata (Johnson and Johnson, 1991) and represents a relatively stable, slowly subsiding platform.

The northern part of the Paradox basin, which contains as much as about 2,400–2,700 m of Pennsylvanian and lower Wolfcampian strata (Mallory, 1972; Rascoe and Baars, 1972), also is in the Uinta-Piceance region (fig. 2). The northern Paradox basin is bounded on the northeast by the Uncompahgre fault and the ancestral Uncompahgre uplift. Based on borehole, seismic, and structural data, Frahme and Vaughan (1983) and White and Jacobson (1983) showed that the Uncompahgre fault was a reverse fault in the late Paleozoic. Frahme and Vaughan (1983) estimated at least 9.6 km of horizontal displacement and about 6 km of vertical displacement along the northern segment of the Uncompahgre fault. The deepest part of the Paradox basin was adjacent to the Uncompahgre uplift (Mallory, 1972, 1975), and the basin's marked asymmetry is consistent with thrust-induced flexural loading along its northeastern margin. The Uncompahgre uplift most likely plunges and dies out to the northwest in a complex structural zone characterized by thrust and reverse faults (Szabo and Wengerd, 1975; Potter and others, 1991).

The Emery uplift or platform (fig. 2) formed a low-relief shelf on the west flank of the Paradox basin and may (at least in part) represent a flexural bulge or "forebulge" coupled to the overthrusted Uncompahgre uplift. Szabo and Wengerd (1975) suggested that the eastern margin of the Emery shelf is fault controlled (the Emery fault), and Potter and others (1991) described northwest-trending reverse faults on the northern flank of the shelf; however, the vertical displacement on these structures is probably not more than a few thousand feet. Permian strata of the northern Paradox basin postdate this postulated faulting and onlap the eastern flank of the Emery uplift. The northern Paradox basin is similar to the Eagle basin in that intrabasinal faults and lineaments had significant effects on depositional patterns and local rates of subsidence (for example, Szabo and Wengerd, 1975; Stevenson and Baars, 1986).

The Emery uplift is bounded on the west by a structurally complex zone characterized by Cretaceous thrust faults (Sevier orogeny) and Neogene extensional faults (Basin and Range). Pennsylvanian and Wolfcampian strata in this province are considered part of the Callville shelf depositional province (fig. 2). Callville shelf deposits probably onlapped the western flank of the low-relief Emery uplift, but this relationship is not clear because of the lack of outcrops and boreholes and because upper Paleozoic strata in this area apparently have been tectonically displaced 50–100 km eastward from their depositional site (Levy and Christie-Blick, 1989, fig. 3).

The southeastern part of the Oquirrh basin is in the northwestern part of the Uinta-Piceance basin region (fig. 2) and contains as much as 7,100 m of Pennsylvanian and Wolfcampian strata (Tooker and Roberts, 1970). Basinal strata have been disrupted by Cretaceous thrust faults and Neogene extensional faults. Levy and Christie-Blick (1989, fig. 3) suggested net eastward tectonic transport of 30–100 km for Oquirrh basin strata in the Uinta-Piceance region. Alternatively, Jordan and Douglas (1980) suggested that, in the area west of the Wasatch fault (the easternmost fault in the extensional Basin and Range province), the younger extensional deformation may have fortuitously cancelled out older Sevier shortening such that late Paleozoic spatial relationships remain relatively intact. Jordan and Douglas (1980) suggested that the Oquirrh basin was a northwest-trending trough that had its maximum topographic relief in the Late Pennsylvanian and Early Permian. The bounding faults of this trough are apparently masked by younger sedimentary rocks and structural features and have not been positively identified. Picha (1986) suggested, however, that the Leamington lineament represents the southern boundary of the Oquirrh basin, and Jordan and Douglas (1980) and Bryant and Nichols (1986) stated that the "Oquirrh-Uinta arch" partly defined the northern margin of the deepest part of the basin. The nature of the boundaries between the Oquirrh basin and the Paradox basin and the Wyoming shelf to the southeast and east are obscured by structural complexities and (or) burial under thick Mesozoic and Cenozoic strata.

In summary, the Uinta-Piceance basin region is characterized by a complex mix of compressional and extensional tectonic styles from the Early Pennsylvanian through the Early Permian. This mix of structural styles is most consistent with regional strike-slip deformation (Christie-Blick and Biddle, 1985), probably driven by interactions along distant continental margins. The relative importance of each continental margin as a driving force for intraplate deformation is explored from the perspective of subsidence analysis later in this paper.

LATE PALEOZOIC CLIMATE, EUSTASY, AND CYCLICITY

The late Paleozoic was the time of the last major pre-Quaternary glaciation (Crowell, 1978). Initiation of this glacial age is generally correlated with the closure of several ocean basins and the resultant amalgamation of continental landmasses in the Southern Hemisphere to form the western half of Pangea (Scotese and McKerrow, 1990). During this period, North America was rotated clockwise about 30°–40° from its present orientation, and the Uinta-Piceance basin region was at a latitude ranging from about 0°–5° N. in the Early Pennsylvanian to about 15° N. in the Early Permian (Scotese and McKerrow, 1990). Extensive eolianites and evaporites in upper Paleozoic deposits of the Uinta-Piceance region (see later discussion) indicate an arid climate. Paleo-

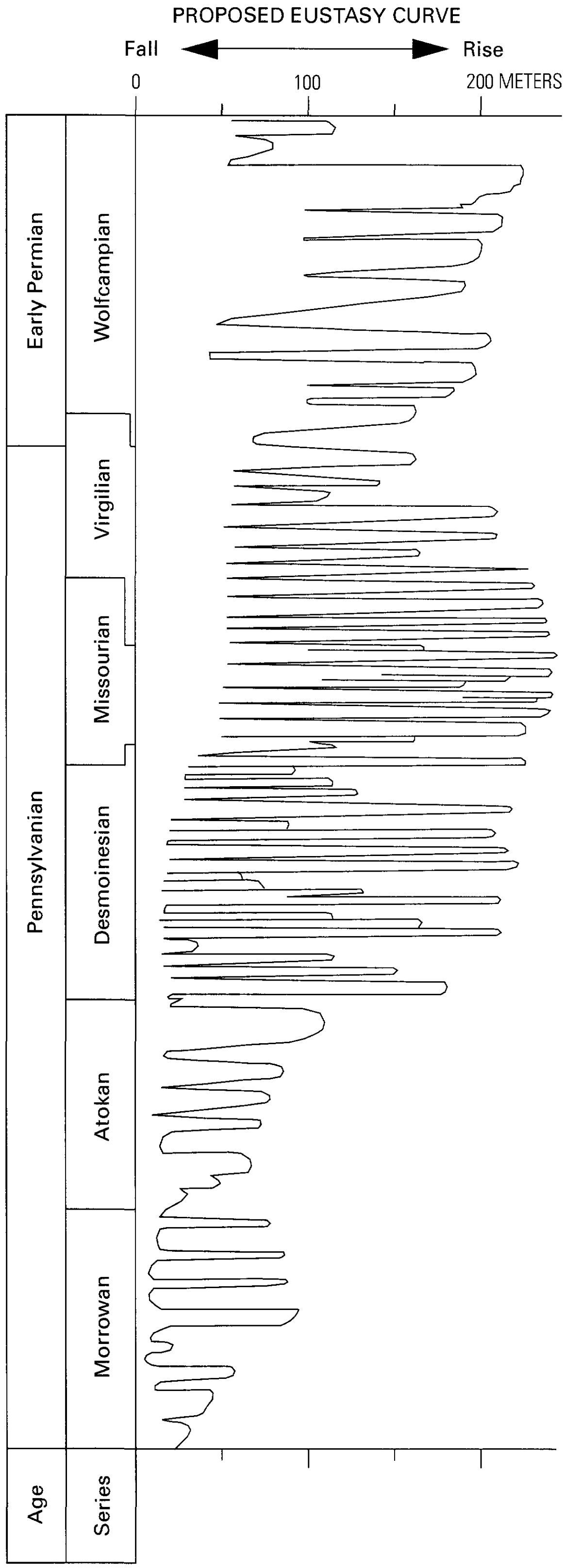

Figure 4. Inferred sea-level curve of Ross and Ross (1987) for the Pennsylvanian and Early Permian.

wind directions were primarily to the west (to the south in present-day coordinates) (Parrish and Peterson, 1988).

Using the Quaternary as an analogue, numerous authors have suggested that the late Paleozoic ice sheets expanded and contracted at fairly regular time intervals, probably forced by Milankovitch orbital parameters (for example, Heckel, 1986). As in the Quaternary, cyclic eustatic and climatic changes are the inferred effects of these ice-sheet fluctuations. Late Paleozoic sea-level curves (for example, Ross and Ross, 1987) (fig. 4) are currently being refined. Sea-level changes are generally interpreted as asymmetric (similar to Quaternary analogues), characterized by rapid transgressions (rapid melting of ice sheets) and slow regressions (slow buildup of ice sheets). Ross and Ross (1987) estimated that the magnitude of Pennsylvanian cyclic eustatic changes ranged from tens of meters to almost 200 m; Kerr (1989) suggested an average sea-level fluctuation of 90 m. Coastal onlap curves (Ross and Ross, 1988, fig. 9) indicate that eustatic fluctuations were superimposed on a significant global sea-level rise from the Morrowan to the end of the Desmoinesian; the magnitude of coastal onlap remained approximately the same in the Missourian, Virgilian, and Wolfcampian.

Deposition during late Paleozoic transgressions and regressions produced sedimentary "cycles" or depositional sequences that have been recognized and correlated across basins and globally (for example, Wanless and Shepard, 1936; Heckel, 1986; Ross and Ross, 1987, 1988). Many of these depositional sequences are not "cyclic" in the strict sense in that preferred ordering of lithofacies may not be present (for example, Konopka and Dott, 1982; Driese and Dott, 1984). This lack of regular ordering within a transgressive-regressive sequence can reflect the variable effects of tectonic activity, sediment supply, basin geometry, and other factors. In this report, we use the term "cycle" in the more general way to describe a transgressive-regressive depositional sequence.

Heckel (1986) estimated that the periodicity of late Paleozoic sedimentary cycles or depositional sequences in the North American midcontinent outcrop belt was approximately 235,000–400,000 years, matching the Milankovitch eccentricity parameter; however, based on a synthesis of 19 studies of late Paleozoic cycles, Kerr (1989) estimated a typical periodicity of 1.3 million years, significantly longer than inferred late Cenozoic eccentricity, and suggested that a longer Milankovitch parameter may have been active in the late Paleozoic. Ross and Ross (1988) suggested that late Paleozoic transgressive-regressive sequences averaged about 2 million years (range of 1.2–4 m.y.) but also noted that these sequences could include four or five of Heckel's (1986) cyclothems. Klein (1990) however, maintained that use of a poorly calibrated time scale by Heckel (1986), and by inference, Ross and Ross (1988) and Kerr (1989) resulted in an overestimation (by a factor of 1.3–1.8) of cycle or sequence periodicity. Klein (1990) challenged the Milankovitch hypothesis and suggested that eustatic changes may have been controlled by other processes including tectonism and large-scale meltwater discharge. Klein (1990) did not discuss a mechanism for repetition of these inferred alternative cycle-controlling mechanisms. Until such a mechanism has been described and justified, Milankovitch orbital parameters of unknown periodicity provide the most compelling rationale for the repetitive late Paleozoic eustatic and climatic changes. Figure 3 shows both the time scale of Haq and Van Eysinga (1987) (which is similar to the scale used by Heckel, 1986) and that proposed by Klein (1990). The Klein (1990) calibrations have been challenged by Langenheim (1991) and Heckel (1991).

PALEOGEOGRAPHY OF THE UINTA-PICEANCE BASIN REGION

Late Paleozoic transgressive-regressive cycles or sequences, primarily reflecting the eustatic controls discussed above, have been reported from most of the well-described stratigraphic units in the Uinta-Piceance basin region (Hite and Buckner, 1981; Driese and Dott, 1984; Johnson and others, 1988; Schenk, 1989; Karachewski, 1992) (fig. 5). Because much of the basinal topography in the Uinta-Piceance region was at or near sea level in the late Paleozoic, inferred eustatic fluctuations significantly affected the location of the marine-nonmarine interface and other depositional patterns. We attempt to show the changes in these depositional patterns on four sets of maps (figs. 6–9) representing deposition during maximum transgression and regression. In one case (late Atokan–Desmoinesian regressions map; fig 7*A*), we also show the limits of an evaporite facies in the Eagle basin that was deposited during but not at the peak of regressions because of the facies' importance in understanding regional depositional patterns. By presenting paired maps, our report differs from previous efforts at regional paleogeographic reconstruction (for example, Hallgarth 1967; Mallory, 1972, 1975; Rascoe and Baars, 1972; Chronic, 1979; Welsh and Bissell, 1979; DeVoto and others, 1986), which generally show the dominant lithofacies or depositional environment in a particular area. This dominant lithofacies or environment may represent deposition during a transgression, a regression, or an intermediate stage.

The time intervals selected for the maps (Morrowan and early Atokan, late Atokan and Desmoinesian, Missourian and Virgilian, early Wolfcampian) coincide with inferred major changes in regional tectonic and (or) depositional patterns. We emphasize that the maps are preliminary models, inhibited by the general lack of outcrop of late Paleozoic rocks (fig. 1) and (or) the lack of detailed sedimentologic studies. We relied extensively on geophysical or lithologic (AmStrat) borehole logs (fig. 1, appendix) for reconstruction of paleogeography in northwesternmost Colorado and the Paradox basin–Emery uplift area; however, these borehole logs generally lack well-constrained age data and systematic lithologic descriptions. We show deposition on the maps in areas where Pennsylvanian rocks are no longer present if there is compelling evidence that the rocks were originally deposited and later eroded. Essentially no data are available from the subsurface of the Paleogene Uinta basin, and the inferred facies patterns in this important area where the Paradox basin merges with the Wyoming shelf and the Oquirrh basin are particularly speculative. Finally, due to gross uncertainties (Jordan and Douglas, 1980; Levy and Christie-Blick, 1989), we did not attempt to palinspastically restore strata of the Oquirrh basin and Callville shelf to their depositional sites (pre-Mesozoic shortening and Cenozoic extension). Future studies will undoubtedly refine estimates of the amounts of tectonic transport and make accurate palinspastic restorations possible.

Pennsylvanian and Early Permian subsidence in four depositional provinces in the Uinta-Piceance basin region is shown in figures 10 and 11. These diagrams are based on considerable imprecise data involving the ages of stratigraphic units and their boundaries, the diagenetic and compaction histories of most units, and calibration of the geologic time scale. Additionally, thicknesses vary significantly for different units within the four sedimentary provinces. The diagrams should therefore be regarded as approximations that illustrate general trends. The ages of units used in the analysis are shown on figure 3. The time scale of Haq and Van Eysinga (1985) was used for all analyses; however, figure 10*A* also shows the data plotted using the time scale proposed by Klein (1990). Use of different time scales modifies the shapes of the curves but not their general trends.

Sub-Pennsylvanian Geology

Morrowan strata unconformably overlie Mississippian rocks throughout most of the Uinta-Piceance region (Mallory, 1972, p. 114; Johnson and Johnson, 1991). In the Eagle basin and most of northwestern Colorado (fig. 4), Morrowan strata generally overlie the Lower Mississippian Leadville or Madison Limestone. Pennsylvanian strata (including the latest Mississippian (?) to early Middle Pennsylvanian Molas Formation) overlie the Leadville Limestone, the Doughnut

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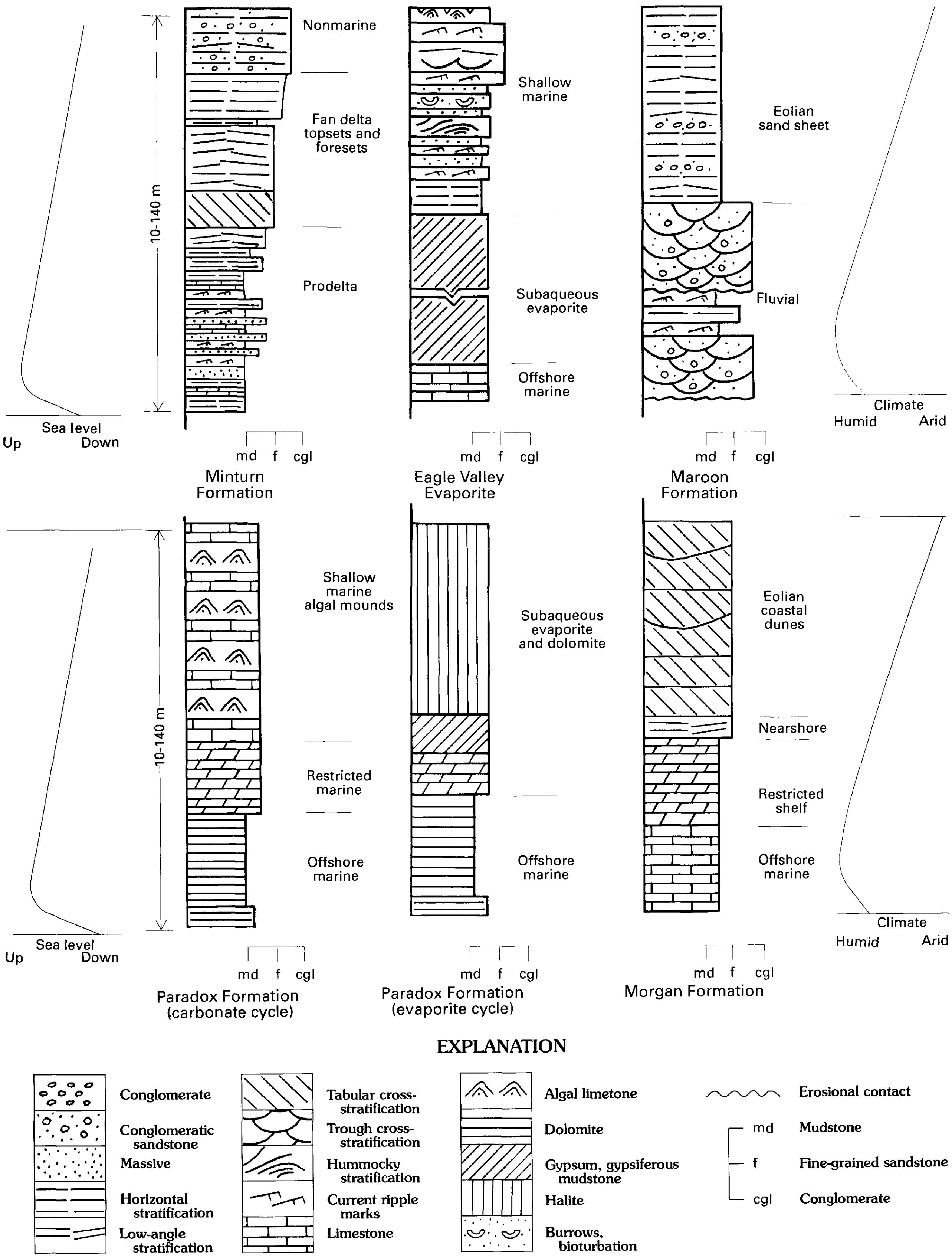
Figure 5 (facing page). Schematic Pennsylvanian depositional sequences in the Uinta-Piceance basin region. Examples from the Minturn and Maroon Formations and the Eagle Valley Evaporite are modified from Johnson and others (1988), Schenk (1989), and Karachewski (1992), from the Morgan Formation modified from Driese and Dott (1984, fig. 20), and from the Paradox Formation modified from Hite and Buckner (1981, fig. 7).



Sea level
Up
Down
10-140 m
Nonmarine
Fan delta topsets and foresets
Prodelta
md f cgl
Minturn Formation
Shallow marine
Subaqueous evaporite
Offshore marine
md f cgl
Eagle Valley Evaporite
Eolian sand sheet
Fluvial
md f cgl
Maroon Formation
Climate
Humid
Arid
Sea level
Up
Down
10-140 m
Shallow marine algal mounds
Restricted marine
Offshore marine
md f cgl
Paradox Formation (carbonate cycle)
Subaqueous evaporite and dolomite
Offshore marine
md f cgl
Paradox Formation (evaporite cycle)
Eolian coastal dunes
Nearshore
Restricted shelf
Offshore marine
md f cgl
Morgan Formation
Climate
Humid
Arid
EXPLANATION
Conglomerate
Conglomeratic sandstone
Massive
Horizontal stratification
Low-angle stratification
Tabular cross-stratification
Trough cross-stratification
Hummocky stratification
Current ripple marks
Limestone
Algal limetone
Dolomite
Gypsum, gypsiferous mudstone
Halite
Burrows, bioturbation
Erosional contact
md
f
cgl
Mudstone
Fine-grained sandstone
Conglomerate

Formation, or the Meramecian Redwall Limestone in the northern Paradox basin area, and the Redwall Limestone underlies Wolfcampian strata on the Emery uplift (Mallory, 1972; Rascoe and Baars, 1972; Welsh and Bissell, 1979).

The contact between the Upper Mississippian Doughnut Formation and the Lower Pennsylvanian Round Valley Limestone in the Wyoming shelf area is problematic. Welsh and Bissell (1979, p. Y21) suggested that "sedimentation was continuous or only slightly interrupted from Chesterian (latest Mississippian) to Morrowan time. The Round Valley Limestone***represents a marine invasion over the deltaic and estuarine environments of the Chesterian Doughnut Formation." Near the Utah-Colorado border, however, some post-Doughnut, pre-Morrowan erosion clearly occurred because the Morgan Formation (in part laterally equivalent to the Round Valley) rests unconformably on the Madison Limestone and the Doughnut Formation is missing (for example, borehole 39, fig. 1). This unconformity may pass westward into a conformity along the Wyoming shelf. To the west, in the Oquirrh basin and on the Callville shelf, Morrowan strata are inferred to conformably overlie Chesterian rocks (Brill, 1963; Tooker and Roberts, 1970). Isopach maps (Poole, 1974; Poole and Sandberg, 1977; Skipp, 1979) indicate that the Oquirrh basin and Wyoming shelf began to subside as discrete tectonic elements in the Late Mississippian, significantly before subsidence began in the Eagle and Paradox basins.

Early to Early Middle Pennsylvanian—Morrowan and Early Atokan

Within the Pennsylvanian and Wolfcampian strata of the Uinta-Piceance region, depositional cycles or sequences are least recognized in the generally thin sections of Morrowan and lower Atokan strata (fig. 6). This relative lack of recognition probably reflects several factors, including poor outcrops, little recent detailed sedimentologic analysis, slow subsidence rates (figs. 10, 11), and eustatic fluctuations of comparatively lesser magnitude. Ross and Ross (1987, 1988) showed 10 significant eustatic fluctuations in the Morrowan and lower half of the Atokan that had inferred magnitudes of from 20 to about 80 m, about half the size of the magnitude inferred for late Atokan through Early Permian eustatic fluctuations (fig. 4).

Brill (1963) described cycles from a 265-m-thick section of the Callville Limestone on the Callville shelf (Route 148 southwest section, fig. 1). These cycles consist of a lower fine-grained sandstone overlain by an upper thin-bedded limestone. Farther south in the Cove Fort area (fig. 1), Morrowan strata (255 m thick) are mainly dolomitic limestone but include some sandstone and oolitic limestone. Brill (1963) suggested that cyclic sedimentation in this area reflects repeated transgression (carbonate platform deposition) and regression (shelf sand deposition) over an unstable border between a shelf to the east and a basin to the west.

Brill (1963) noted only a thin Atokan sequence of the Callville Limestone in the Callville shelf area, overlain by an unconformity. Welsh (1976) argued that Desmoinesian strata were deposited across the Callville shelf but subsequently eroded. Brill (1963) described cycles from the Atokan part of the Callville from the Route 148 southwest section (fig. 1) that consist of a basal gray to brown shale overlain by a middle unit of gray to yellowish-gray sandstone and an upper unit of thin-bedded limestone or dolomite. Similar sandstone is crossbedded in the Cove Fort section (fig. 1). Brill did not describe the limestone, but it probably represents deeper facies (transgressive deposits) deposited at the beginnings of cycles, whereas the sandstone forms the upper parts of cycles and was deposited during regressions.

In the Oquirrh Mountains (fig. 1) of the southeastern Oquirrh basin, the Lake Point Limestone and the West Canyon Limestone include about 365–440 m of mainly Morrowan strata. Tooker and Roberts (1970, p. A12) described cycles in the Lake Point Limestone that comprise, in ascending order, shaly limestone, clastic calcareous sandstone, calcitic arenaceous limestone, crossbedded arenaceous limestone, cherty limestone, and bedded chert. They suggested (p. A24) that the West Canyon Limestone, comprising arenaceous limestone, cherty argillaceous limestone, and crossbedded sandstone, is also characterized by cyclic deposition but did not describe a typical cycle. These cycles resemble those described for the Callville shelf and have an inferred similar origin. Alexander (1978) measured and provided detailed petrographic descriptions of a section 321 m thick of similar age and lithology at Cascade Mountain in the central Wasatch range. He did not identify cycles but interpreted a variety of shallow-marine depositional environments that changed in response to variations in sea level.

Tooker and Roberts (1970) and Konopka and Dott (1982) described a significant change in lithology and depositional style in the southeastern Oquirrh basin between Morrowan strata and the more sandstone-rich Atokan to Desmoinesian Butterfield Peaks and Erda Formations. The timing of this change is similar to that noted for the Callville shelf. These changes are shown on the late Atokan to early Desmoinesian maps (fig. 7) but were probably initiated in the early to middle Atokan.

The Emery uplift, on which Permian rocks unconformably overlie Mississippian strata, bounds the Callville shelf to the east. Strata on the Callville shelf are relatively fine grained, and it is possible that Morrowan and lower Atokan strata were deposited over at least the westernmost part of the Emery uplift (as shown in fig. 6) and subsequently eroded. The area that now includes the eastern part of the Emery uplift, the Paradox basin, and at least part of the ancestral Uncompahgre uplift was probably emergent or partly emergent. Latest Mississippian (?) through early

Atokan deposition in the Paradox basin is recorded by the Molas Formation, mainly a red regolith that formed by weathering of Mississippian limestone. The emergent area is here informally named the "Molas platform."

Morrowan and lower Atokan strata of the Wyoming shelf depositional province are generally less than about 150 m thick (Johnson and Johnson, 1991) and include the Round Valley Limestone and the lowest part of the Morgan Formation. The Round Valley Limestone crops out along the south and north flanks of the Uinta Mountains and extends east as far as Juniper Mountain (fig. 1) (Upton, 1958). It consists mainly of thin-bedded to massive limestone, commonly dolomitic or cherty (Sadlick, 1955, 1957; Hansen, 1965). Siliciclastic deposits, generally gray or purplish-red mudstone, are typically a minor component but locally (for example, Sols Canyon, fig. 1) comprise as much as 35 percent of Round Valley sections. Huddle and McCann (1947) noted a significant clastic component to the Round Valley Limestone (mapped as the lower member of the Morgan Formation) in the Duchesne River area (fig. 1).

The proportion of clastic material in Morrowan and lower Atokan strata (or strata that overlie Mississippian rocks and are of inferred Morrowan and early Atokan age) increases in the subsurface south and southeast of the Uinta Mountains outcrop belt (for example, boreholes 34–38, fig. 1), in strata assigned to the Morgan Formation. This increase suggests that the "Molas platform" contributed clastic detritus to the Wyoming shelf. The more abundant limestone in this province was probably deposited during transgressions, when clastic detritus was trapped closer to its source. Regressions were probably characterized by lowering of base level and more widespread clastic deposition.

The Morrowan and lower Atokan Belden Formation (as thick as 200 m or more) of the Eagle basin represents deltaic and prodeltaic deposition off the ancestral Front Range. The best exposures of cyclic deltaic deposits are in the Sweetwater Creek area about 30 km northeast of Glenwood Springs (fig. 1). Typical cycles coarsen upward from (1) prodeltaic deposits of gray to black shale, dolomite, and limestone to (2) delta-front deposits of wave-rippled, fine- to coarse-grained sandstone and interbedded dark mudstone. Some cycles are capped by fluvial channel deposits of trough crossbedded, coarse-grained to conglomeratic sandstone. Paleocurrent data ($n=60$, $X=237°$, $\sigma=41°$) indicate that sediment transport was to the southwest. The texture and sedimentology of these deltaic rocks indicate contemporaneous uplift in the ancestral Front Range. Houck (1991) noted a large delta system in the overlying Minturn Formation in the McCoy area (fig. 1), in the same location as the large Belden delta shown in figure 6*A*. She inferred that the large Minturn delta was fed by a major drainage system channeled through a west-trending transfer zone in the Gore fault, the interpretation shown on figure 6.

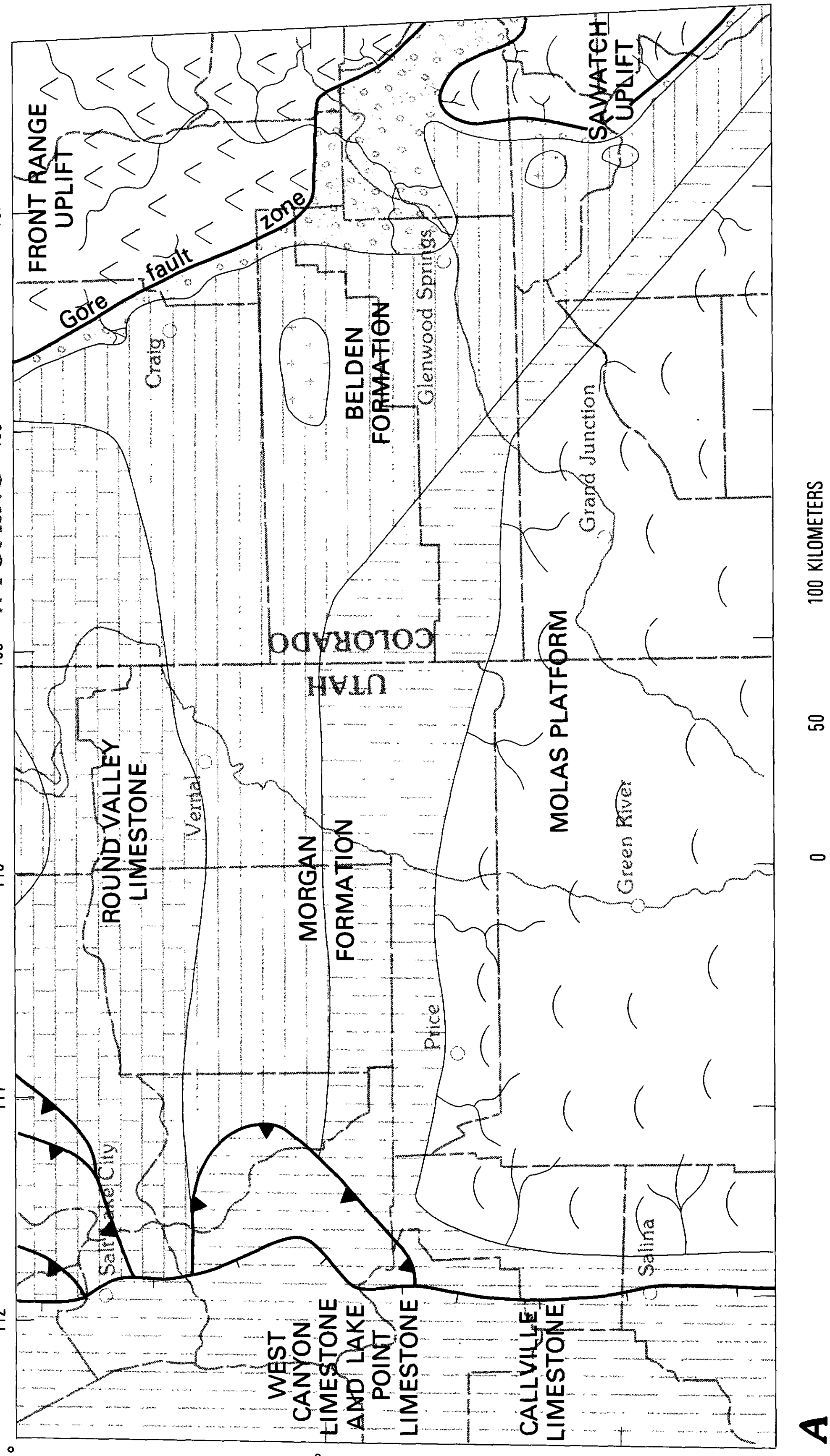
Throughout most of Eagle basin, the Belden Formation consists of interbedded gray limestone and dark-gray, commonly calcareous, mudstone. These mudstones are the inferred regressive deposits and represent maximum progradation of deltas and associated prodelta clastic sediments. During transgressions, clastic sediments were trapped closer to uplifted sources and limestones were deposited over broader areas. Beds of anhydrite and (or) gypsum in the Belden Formation have been found in the subsurface on the north flank of the White River uplift (for example, boreholes 41–43, fig. 1), and in outcrops on the west side of the Sawatch uplift (Freeman, 1972; Bryant, 1979). Interbedded facies (shale and limestone) at each location suggest a basinal setting. We infer that these evaporites precipitated from basinal brines in fault-bounded sub-basins (following DeVoto and others, 1986; Dodge and Bartleson, 1986) during eustatic lowstands.

DeVoto and others (1986) presented evidence from an area southeast of the Uinta-Piceance region that major coarse-grained delta systems in the Belden Formation had a source in the Sawatch uplift. The lack of comparable deltaic deposits on the west side of the Sawatch uplift (Bryant, 1979) in the Uinta-Piceance region suggests either that the uplift had markedly asymmetric relief or that it was significantly narrower than the modern Sawatch Range.

Based on the dominance of limestone and dark mudstone in the Belden Formation in the Crested Butte area (fig. 1), Langenheim (1978) concluded that the ancestral Uncompahgre uplift had not developed in the Morrowan and early Atokan. The "Molas platform" apparently did not shed significant sediment eastward into the Eagle basin. The eastern boundary of this low-relief source terrane was probably west of the Middle and Late Pennsylvanian margin of the Uncompahgre uplift.

Due to a lack of subsurface data, the location of the Morrowan–early Atokan structural boundary between the Front Range and the Eagle basin is not well constrained. North and northwest of the McCoy area (fig. 1), coarse-grained deltaic deposits have not been recognized adjacent to the basin margin, either in outcrops (for example, Sharps, 1962) or in the subsurface (for example, Thomas and others, 1945; Hallgarth, 1959) (boreholes 44, 45, 47, 51, fig. 1). Morrowan and lower Atokan strata are probably not present in borehole 53 (fig. 1), where a thin (120 m) Pennsylvanian section rests on Mississippian rocks. It may be that Morrowan and lower Atokan rocks were never deposited east of borehole 53, as suggested by DeVoto and others (1986, fig. 4). Alternatively, Morrowan and lower Atokan rocks could have been deposited east of borehole 53 and then eroded later in the Pennsylvanian.

Summarizing, Morrowan and lower Atokan strata across most of the Uinta-Piceance basin area consist of alternating fine-grained clastic rocks (inferred regressive deposits) and more abundant limestone (transgressive deposits). Significant Morrowan and early Atokan tectonic uplifts include the ancestral Front Range and Sawatch uplifts.



A
WYOMING
COLORADO
UTAH
FRONT RANGE UPLIFT
Gore fault zone
SAWATCH UPLIFT
BELDEN FORMATION
ROUND VALLEY LIMESTONE
MORGAN FORMATION
MOLAS PLATFORM
WEST CANYON LIMESTONE AND LAKE POINT LIMESTONE
CALLVILLE LIMESTONE
Craig
Glenwood Springs
Grand Junction
Vernal
Green River
Price
Salt Lake City
Salina
107°
108°
109°
110°
111°
112°
41°
40°
39°
0
50
100 KILOMETERS

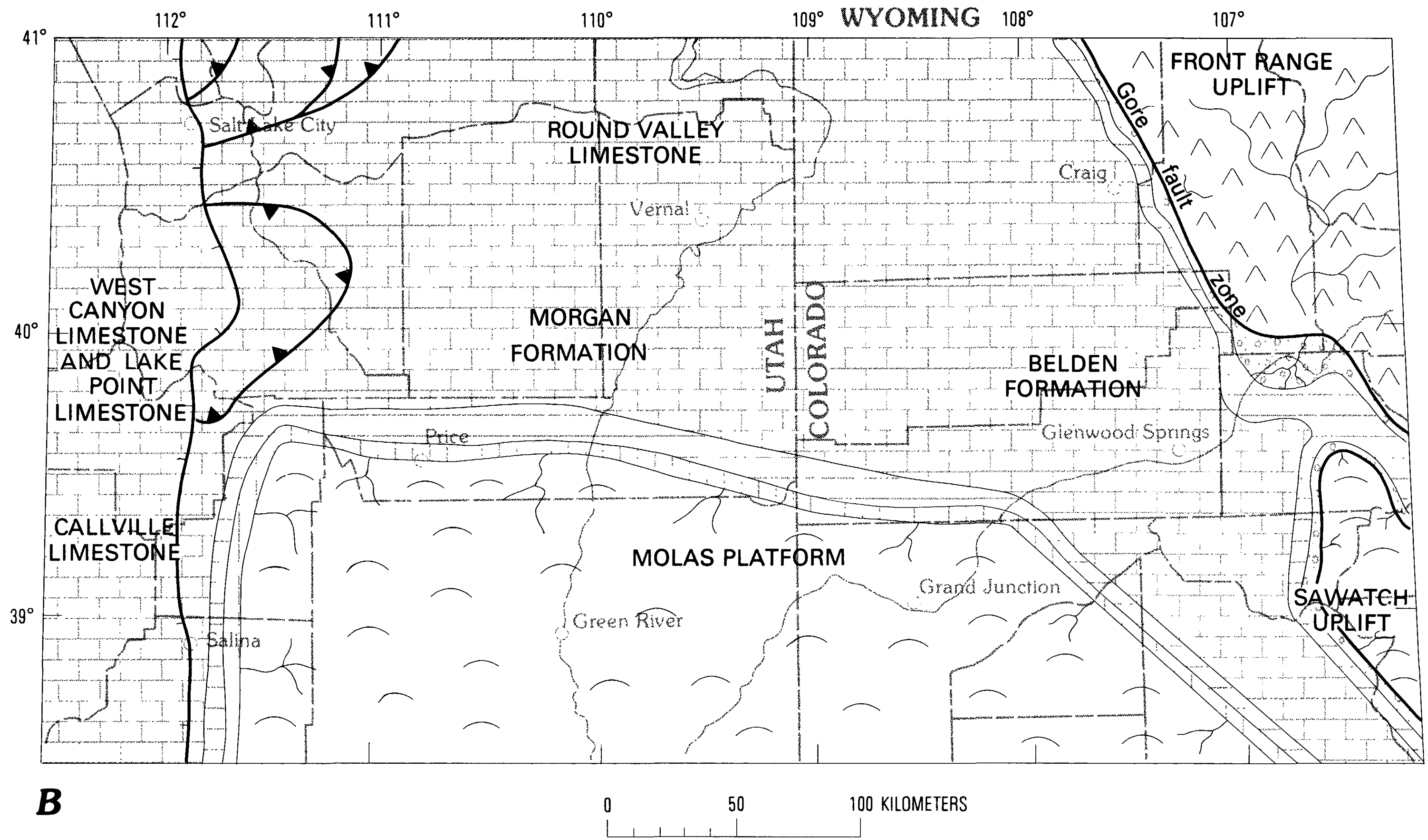
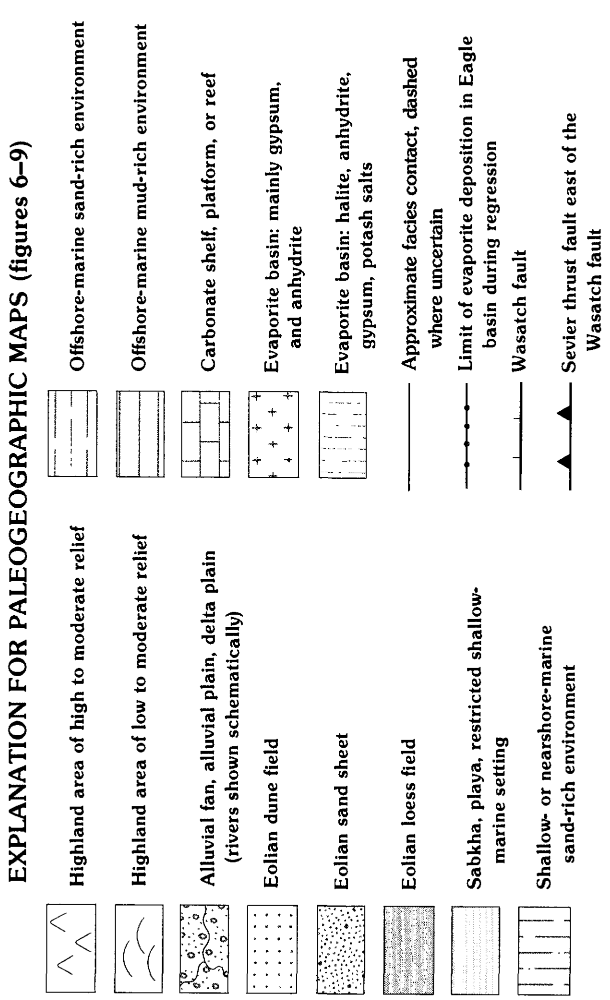


Figure 6. Maps showing the Morrowan to early Atokan paleogeography of the Uinta-Piceance basin region during maximum regression (*A*) and transgression (*B*).



Explanation for figures 6 through 9.

Middle and Late Middle Pennsylvanian—Late Atokan and Desmoinesian

The late Atokan and Desmoinesian paleogeographic maps (fig. 7) highlight three major changes that occurred during the early and (or) middle part of the Atokan in the Uinta-Piceance region: (1) uplift of the ancestral Uncompahgre uplift and the Emery uplift; (2) subsidence of the Paradox basin; and (3) significant progradation of clastic sediments across the Wyoming shelf. Ross and Ross (1987, 1988) recognized 24 major eustatic fluctuations during the last half of the Atokan (2) and Desmoinesian (22) having inferred magnitudes of as much as 200 m (fig. 4). The sea-level curve of Heckel (1986) is quite similar.

As described earlier, upper Atokan and Desmoinesian strata probably have been eroded from the Callville shelf (Welsh, 1976). Accordingly, figure 7 shows the early Atokan depositional pattern (described above) as continuing into the late Atokan and Desmoinesian. Brill (1963) suggested that the increased clastic detritus on the Callville shelf was derived from both the Uncompahgre and Emery uplifts. The Emery uplift was emergent, but preservation of Mississippian rocks below a Permian unconformity indicates that this uplift had minimal relief, probably was not a major sediment source, and may have been onlapped significantly during maximum transgressions.

The Butterfield Peaks and Erda Formations represent Atokan and Desmoinesian deposition in the Oquirrh basin (Tooker and Roberts, 1970). From detailed sedimentologic studies in the Mount Timpanagos area (fig. 1) on the upper plate of the Charleston-Nebo thrust and in the southeastern Oquirrh basin, Konopka and Dott (1982) and Konopka (1985) described lithofacies deposited in environments ranging from normal-marine carbonate shelf to eolian dunes, inferred to represent deposition during maximum transgressions and regressions, respectively. The statistical analysis presented in these studies suggests that these rocks are characterized by several preferred vertical facies transitions but were not cyclic in the strict sense. Suggested factors that might have obscured better development of cycles include (1) rapid, possibly episodic subsidence, (2) rapid deposition of windblown sand and silt, and (3) rapid deposition of carbonate rocks in environments not overwhelmed by sand (Konopka and Dott, 1982). Eolian deposits are much less common in the Oquirrh Group west of the Wasatch fault than they are at Mount Timpanagos. Shoal and emergent carbonate deposits are present at several locations west of the Wasatch fault. These deposits, along with too numerous alternations of sandy and carbonate facies, show that there is not a one-to-one correspondence between lithology and transgressive-regressive depositional patterns.

Tooker and Roberts (1970), Morris and others (1977), and Konopka (1985) described similar lithofacies in the Oquirrh Group in the Oquirrh Mountains and at several other locations west of the Wasatch fault. A number of authors (for example, Nygreen, 1958; Bissell, 1962; Roberts and others, 1965; Tooker and Roberts, 1970) suggested that these strata are cyclic, mainly on the basis of the obvious alternation of siliciclastic and carbonate sediments. Stevens and Armin (1983) described similar Middle Pennsylvanian rocks immediately west of the Uinta-Piceance basin region. They described an idealized transgressive-regressive facies sequence within the carbonate strata, bounded by siliciclastic units that represent the most regressive facies.

Based on extensive paleocurrent measurements, Konopka and Dott (1982) and Konopka (1985) reported that sand transport in the Oquirrh basin during the Atokan and Desmoinesian was to the south. In addition, sandstone deposition was slightly more dominant in the northern and eastern parts of the basin than in the southern and western parts. The sands probably were derived from the craton to the north and northeast and transported to the Oquirrh basin primarily via the Wyoming shelf, and significant sand input from either the Emery or Uncompahgre uplift seems unlikely. Mean composition of siliciclastic grains (2–3 phi) is within the quartz-rich end of the subarkose field, a mature composition for this grain size (Odom and others, 1976). Jordan (1979) and Jordan and Douglas (1980), however, pointed out that, because the Oquirrh basin subarkosic sands (Late Pennsylvanian and Early Permian) and Pennsylvanian subarkosic sands derived from crystalline rocks in the Front Range uplift are compositionally similar, the source of sands in the Oquirrh basin may have been the crystalline rocks of the nearby Uncompahgre uplift. Welsh and Bissell (1979, fig. 8) inferred that a delta system derived from the Uncompahgre uplift fed the Oquirrh basin during the Middle and Late Pennsylvanian, but their hypothesis is difficult to assess because neither outcrops nor subsurface data are available in the critical area between the Uncompahgre uplift and the Oquirrh basin and fluvial-deltaic strata have not been recognized in the Oquirrh basin. Potter and others (1991) documented the presence of a complex structural zone on the northwest flank of the Uncompahgre uplift that might have served to channel Uncompahgre detritus to the northwest. If Uncompahgre-derived sand did enter the southern part of the Oquirrh basin (as shown in fig. 7*A*), then it probably was reworked and transported south onto the Callville shelf.

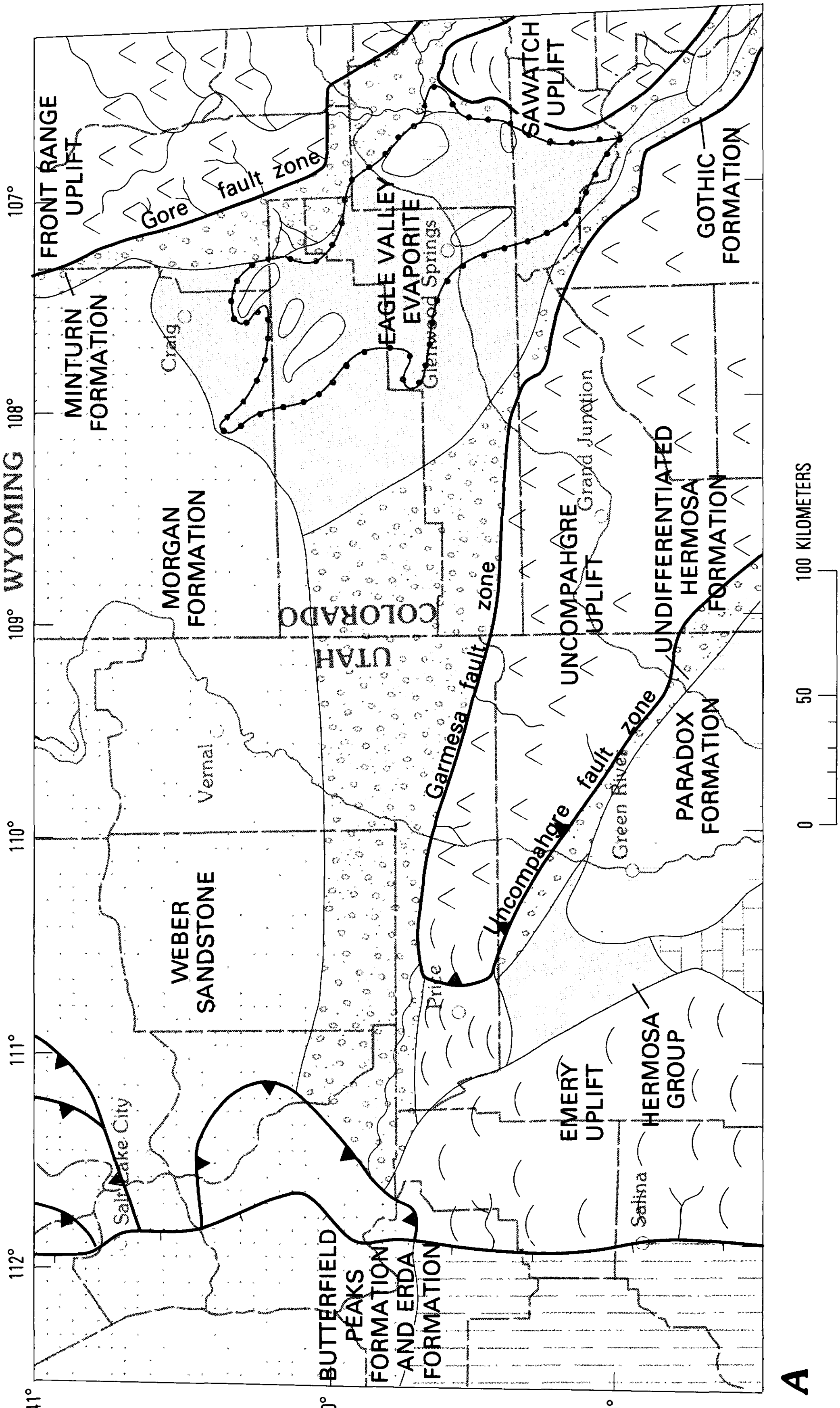
Driese and Dott (1984) described sedimentary cycles in the upper Atokan and Desmoinesian upper member of the Morgan Formation of the Wyoming shelf depositional province (fig. 5). These cycles typically pass upward from (1) carbonate rocks deposited in normal-marine shelf environments (maximum transgression deposits), to (2) carbonate rocks deposited in restricted-shelf and nearshore environments, to (3) sandstones deposited in eolian dunes (maximum regression deposits). Sands were derived from the north, across the Wyoming shelf.

Transgressive-regressive depositional patterns on the Wyoming shelf and in the Oquirrh basin (and by inference, on the Callville shelf) were thus fairly similar. Atokan-Desmoinesian strata on the Wyoming shelf (Morgan Formation and lower part of Weber Sandstone) have a maximum thickness of about 460 m, whereas correlative strata in the Oquirrh basin (the Butterfield Peaks Formation) are as thick as 2,765 m (Tooker and Roberts, 1970). Because sediment supply and subsidence were in balance in the Oquirrh basin, the basin was not characterized by significant negative topographic relief at this time. If these two areas experienced the same number of transgressive-regressive events during this period, then the Oquirrh basin depositional cycles should be much thicker than those that formed on the Wyoming shelf.

The Paradox basin formed as a discrete tectonic element in the Atokan by fragmentation of the "Molas platform." The low-relief Emery uplift (discussed above) and the Uncompahgre uplift (of inferred high relief) formed the western and east-northeastern margins of the Paradox basin, respectively. Erosion of Paleozoic cover rocks from the Uncompahgre uplift exposed Precambrian crystalline rocks, considered to be the main source of clastic detritus in the Paradox basin. As described earlier, seismic data (Frahme and Vaughan, 1983) show that the late Paleozoic boundary between the northern Paradox basin and the Uncompahgre uplift was a reverse fault, and basin asymmetry is inferred to reflect flexural loading along the northeastern margin of the basin.

Mallory (1972, figs. 6, 7) showed a maximum thickness of about 1,700 m for the Atokan and Desmoinesian in the northern Paradox basin, including strata of the Pinkerton Trail and Paradox Formations of the Hermosa Group and undivided rocks of the Hermosa Formation. The Atokan to lower Desmoinesian Pinkerton Trail Formation is the lowest unit and consists of mixed carbonate and fine-grained clastic rocks. The Pinkerton Trail is thickest on the western margin of the basin and thins toward the center of the basin, where it intertongues with and is overlain by the Paradox Formation, which consists of evaporites, black shale, and limestone (Hite and Liming, 1972). The undivided rocks of the Hermosa Formation crop out on the northeastern margin of the basin and are coarse-grained clastic rocks derived from the Uncompahgre uplift.

Atokan and Desmoinesian sedimentary cycles in the Paradox basin are well described (fig. 5) (Peterson and Hite, 1969; Hite, 1970; Hite and Buckner, 1981). Hite and Liming (1972, p. 134–135) identified 29 evaporite cycles on a Paradox basin cross section that includes boreholes 7 and 13 (fig. 1). These basinal evaporite cycles ideally consist of highly calcareous, organic-rich black shale (inferred maximum transgression deposits), anhydrite, dolomite and silty dolomite (intermediate sea-level stand), and halite and potash salt deposits (maximum regression). Some halite beds exhibit evidence of subaerial exposure and erosion. Hite and Cater (1972) showed the limits of evaporite occurrence. On the western margin of the basin (for example, boreholes 7 and 13, fig. 1), Hite and Buckner (1981) described cycles that


FRONT RANGE UPLIFT
Gore fault zone
SAWATCH UPLIFT
GOTHIC FORMATION
MINTURN FORMATION
EAGLE VALLEY EVAPORITE
Glenwood Springs
Craig
WYOMING
MORGAN FORMATION
COLORADO
UTAH
Grand Junction
UNCOMPAHGRE UPLIFT
UNDIFFERENTIATED HERMOSA FORMATION
Garmesa fault zone
Uncompahgre fault zone
PARADOX FORMATION
Green River
Vernal
WEBER SANDSTONE
Price
EMERY UPLIFT
HERMOSA GROUP
Salina
Salt Lake City
BUTTERFIELD PEAKS FORMATION AND ERDA FORMATION
107°
108°
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A

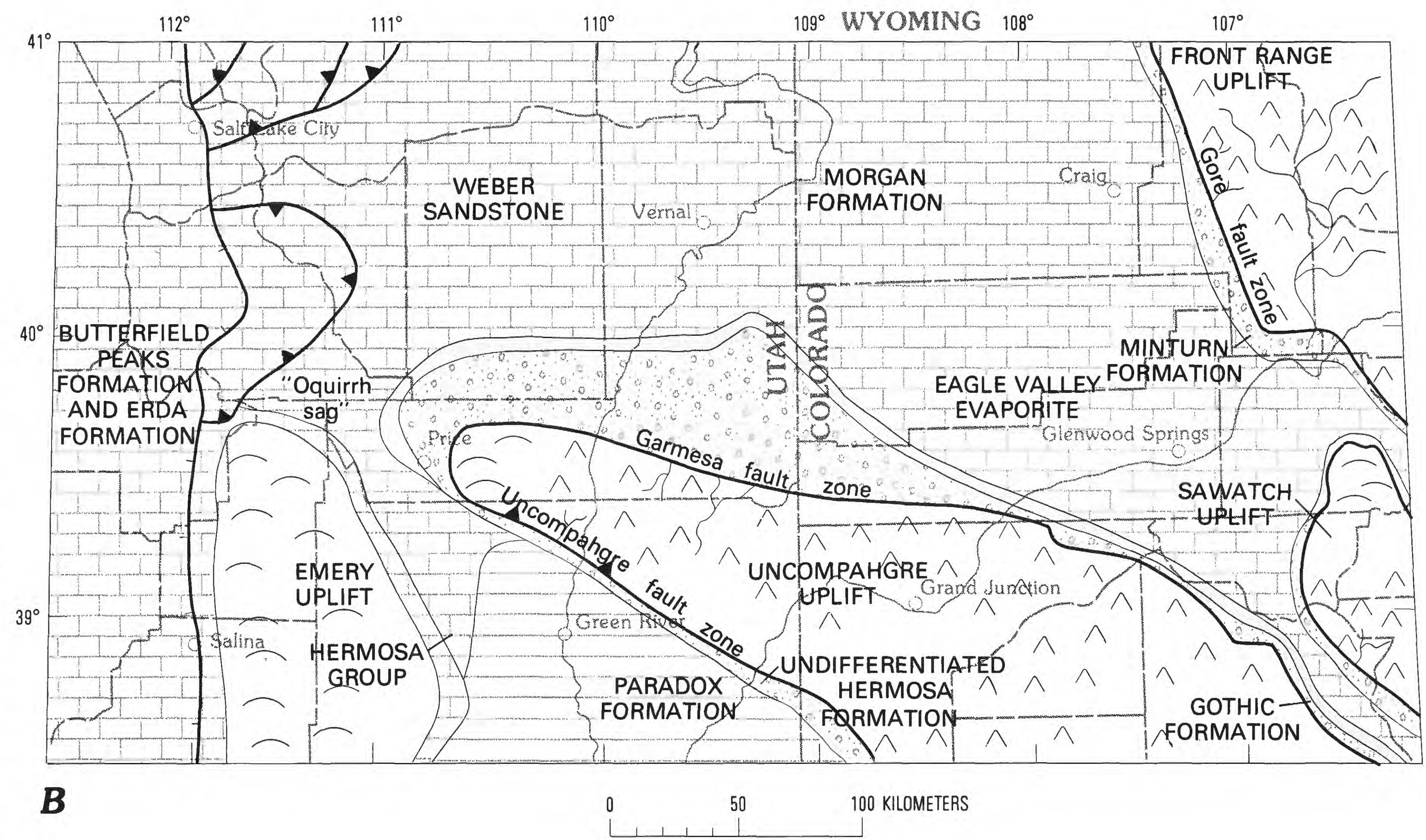


Figure 7 (above and facing page). Maps showing the late Atokan to Desmoinesian paleogeography of the Uinta-Piceance basin region during maximum regression (*A*) and transgression (*B*). See explanation on page CC14.

ideally consist of calcareous black shale (inferred maximum transgression deposits), dolomite and siltstone (intermediate sea level), and algal and fossiliferous limestone (maximum regression). Strata north of the evaporite limit in the northernmost Paradox basin (for example, boreholes 16, 18, 26, fig. 1) consist mainly of interbedded fossiliferous limestone (inferred transgressive deposits), and variegated fine-grained clastic deposits of inferred sabkha or playa origin (inferred regression deposits).

It should be noted that the basinal and western-basin-margin cycles of the Paradox described above are anomalous in comparison to other regionally correlative cycles in that the most clastic rich intervals have been assigned to transgressive phases (that is, raised base level), whereas clastic sediments elsewhere are inferred to have been trapped in basin-margin settings. Hite and Buckner (1981) suggested that the clastic component in the Paradox basin reflects reworking of basin-margin clastic sediments by transgressing seas. Perhaps a more suitable explanation involves sediment accumulation rates. The thick halite intervals in the cycles may have deposited at rates of 5 cm per year or more (O.B. Raup, U.S. Geological Survey, oral commun., 1991), whereas the highly calcareous, organic-rich, black shales probably accumulated at rates of a few millimeters per year (Hite and Buckner, 1981) or much less. Given these rates, it may be that input of clastic sediment to the basin actually increased during regressions, but that this increase is masked by the great volume of the rapidly accumulating evaporites.

Hite and Buckner (1981) postulated that the evaporites were deposited subaqueously in a deep basin through precipitation from a concentrated brine, an interpretation that requires a basin with restricted circulation. The Emery uplift was emergent and provided a partial barrier to circulation. Szabo and Wengerd (1975, fig. 11) showed several possible inlets to the Paradox basin, including (1) the "Oquirrh sag," a presumed connection with the Oquirrh basin in the structurally complex zone between the northwestern end of the Uncompahgre uplift and the northern Emery uplift (Potter and others, 1991), and (2) the "Fremont sag," a possible east-west connection between the Paradox basin and the Callville shelf, to the south of the Uinta-Piceance region. Given the nonmarine to marginal-marine regressive facies described above for the Oquirrh basin and Callville shelf, it is unlikely that either inlet was active during maximum regressions. The presence of fossiliferous limestone in the northernmost Paradox basin, however, requires normal or near-normal marine waters in the basin during transgressions, and this water must have entered the basin without mixing with the evaporitic brines. This nonmixing requirement is more easily met for the Oquirrh sag, and we therefore show it as an inlet into the Paradox basin during maximum transgressions. Boreholes in the area of the Oquirrh sag (for example, boreholes 24, 25, 27, 28, fig. 1) contain anomalously thin sections (less than 220 m) of inferred but undated Pennsylvanian strata and do not provide sufficient data to test the hypothesis.

Tectonic activity was a major control on evaporite deposition in the Paradox basin. Based on apparent rates of subsidence in the Paradox basin (figs. 10, 11), the Uncompahgre uplift probably experienced its highest rates of overthrusting, and extended farthest to the northwest, during the Desmoinesian. Decreased activity on the Uncompahgre would have resulted in decreased sedimentation rates (and accommodation space for evaporites), greater progradation of clastic sediments from basin margins, and widening of the Oquirrh sag, all effects detrimental to evaporite deposition and preservation.

The boundary between the Eagle basin and Wyoming shelf depositional provinces is gradational north of the Uncompahgre uplift near the Colorado-Utah State line. In general, varicolored clastic rocks become more fine grained and are interbedded with greater proportions of limestone with increasing distance from the Uncompahgre uplift (boreholes 33–39, fig. 1). Anomalous limestone-poor strata in borehole 35 indicate lateral lithologic variability and the possible presence of a north-flowing fluvial-deltaic system. During regressions, alluvial clastic sediments derived from the Uncompahgre probably merged northward with south-prograding eolian deposits. The greater width of the belt of Uncompahgre-derived clastic sediments north of the uplift in Utah than in the Paradox and Eagle basins probably reflects the lower subsidence rates characteristic of the Wyoming shelf depositional province. In more rapidly subsiding areas, Uncompahgre-derived clastic sediments are inferred to have been trapped closer to basin margins.

During regressions, alluvial clastic sediments derived from the northern Front Range uplift probably interfingered with south-prograding dunes in the northeastern Eagle basin, but there is little subsurface data to document this relationship. The southern limit of the Wyoming shelf dune field probably was in the Juniper Mountain area (fig. 1). Thick beds of white and red sandstone, presumably eolian, are present in the upper part of the Morgan Formation in boreholes 48 and 51 (fig. 1) and at Juniper Mountain (Upton, 1958). Thick beds of well-sorted sandstone are almost absent farther southeast in boreholes 47 and 49.

Upper Atokan and Desmoinesian strata of the Eagle basin include the Minturn and Gothic Formations on the eastern and western margins of the basin, respectively, and the Eagle Valley Evaporite in the central part of the basin. Karachewski (1992) described highly variable shoaling-upward sequences (fig. 5) in the Minturn Formation in the Minturn area (fig. 1). Maximum regression deposits are coarse alluvial and fan-delta clastic rocks; maximum transgression deposits are offshore marine mudstone and biostromal limestone. During transgressions, limestone was deposited within a few kilometers of the Gore fault (Houck, 1991; Karachewski, 1992) (borehole 46, fig. 1). Karachewski (1992) inferred that relief was significant on the Front Range uplift, the source for the Minturn Formation, and minor on the Sawatch uplift, which he considered an influence on depositional patterns but not a significant sediment source. Houck (1991) inferred that a large deltaic system in the McCoy area (fig. 1) was channelled through a west-trending transfer zone in the Gore fault.

Leighton (1986) and Levorson (1986) reported shoaling-upward sequences similar to those in the Minturn Formation from the Gothic Formation in the Crested Butte and Crystal River areas (fig. 1), respectively. Each concluded that both the Uncompahgre and Sawatch uplifts were significant sediment source terranes and that relief was greater on the Sawatch. Leighton inferred that the main input of marine waters into the "Gothic trough" was from the south.

In the central part of Eagle basin (Eagle Valley location of fig. 1), Schenk (1989) described shoaling-upward cycles in the Eagle Valley Evaporite (fig. 5). Micritic carbonates are inferred maximum transgression deposits. Because Desmoinesian subsidence rates were greater in the Eagle basin than on the Wyoming shelf, sea water was apparently trapped in the Eagle basin during regressions and gypsum was precipitated from an evaporative brine for much of the duration of the regressive phase. Although gypsum is widespread over the central part of Eagle basin, halite is present in only a few discrete areas (for example, DeVoto and others, 1986, p. 43; Dodge and Bartleson, 1986, p. 117). These areas of thick halite deposition have been interpreted as topographically low areas produced by syndepositional faulting (DeVoto and others, 1986; Dodge and Bartleson, 1986). During maximum regressions, shallow marine to nonmarine clastic sediments prograded over evaporitic deposits, and access to normal marine water was completely cut off to the north by Morgan eolianites. Consistent with the inferred paleogeography, the lack of a marine fauna in clastic rocks of the Eagle Valley Evaporite suggests that lowstands (regressive phases) were characterized by lacustrinelike circulation (Schenk, 1989). Schenk (1989) noted that the geometry and facies of clastic wedges at the tops of cycles in the Eagle Valley Evaporite indicate progradation from both the Uncompahgre and Front Range uplifts.

The location of the northern part of the Gore fault zone may have moved eastward from its Morrowan to early Atokan position. The very thin Pennsylvanian section in boreholes 52 and 53 (fig. 1), closest to the Front Range uplift, is consistent with a complex history of basin-margin faulting.

Summarizing, the late Atokan to Desmoinesian history of the Uinta-Piceance region is characterized by increased tectonic activity reflected in higher rates of basin subsidence and in continued or initial uplift and unroofing of the Front Range, Sawatch, and Uncompahgre uplifts. The combined effects of tectonism and eustasy led to restricted circulation in the Eagle and Paradox basins in which evaporite deposition was significant. Deposition during regressions was also characterized by significant progradation of eolian sands across the Wyoming shelf. Deposition of limestone dominated during transgressions.

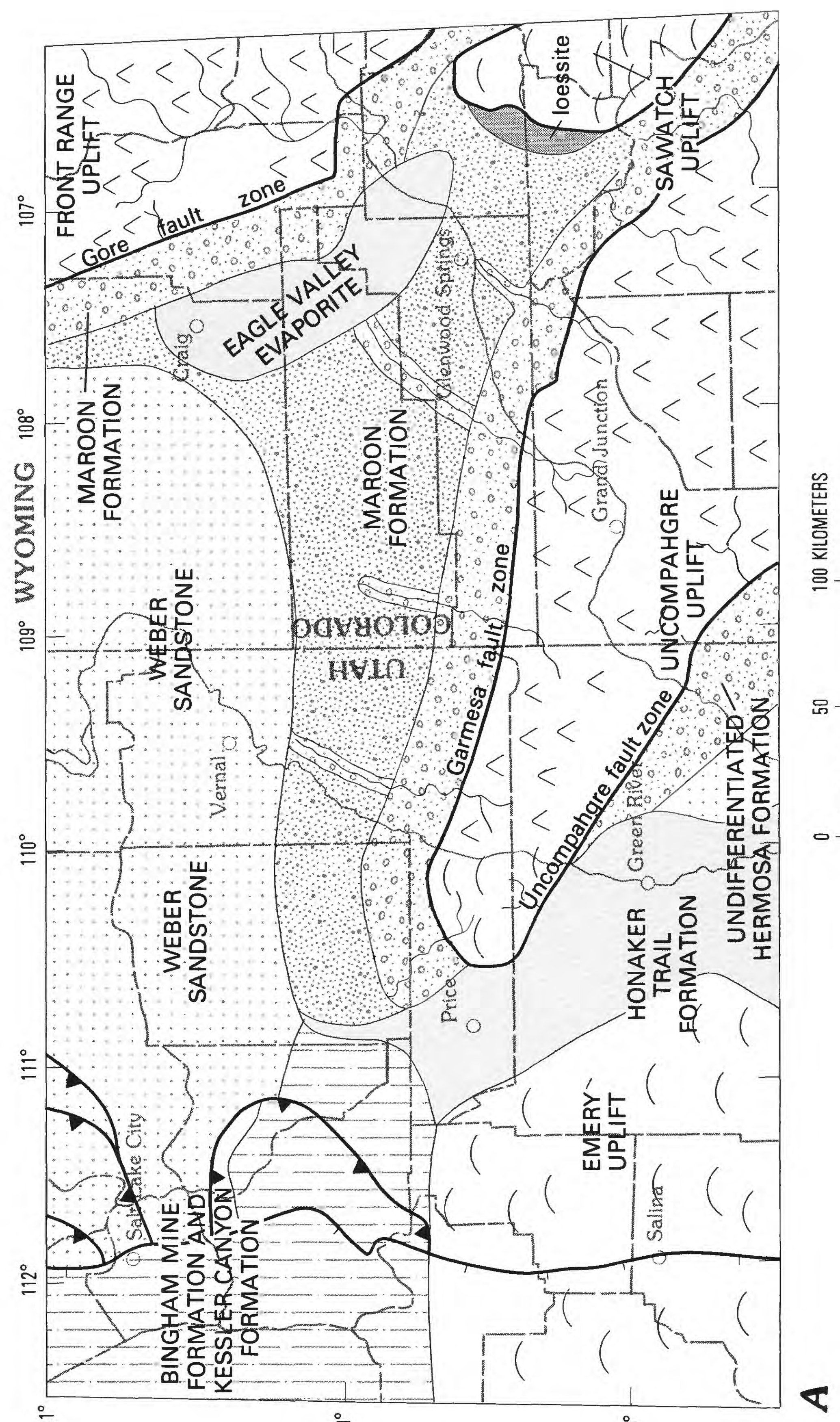
Late Pennsylvanian—Missourian and Virgilian

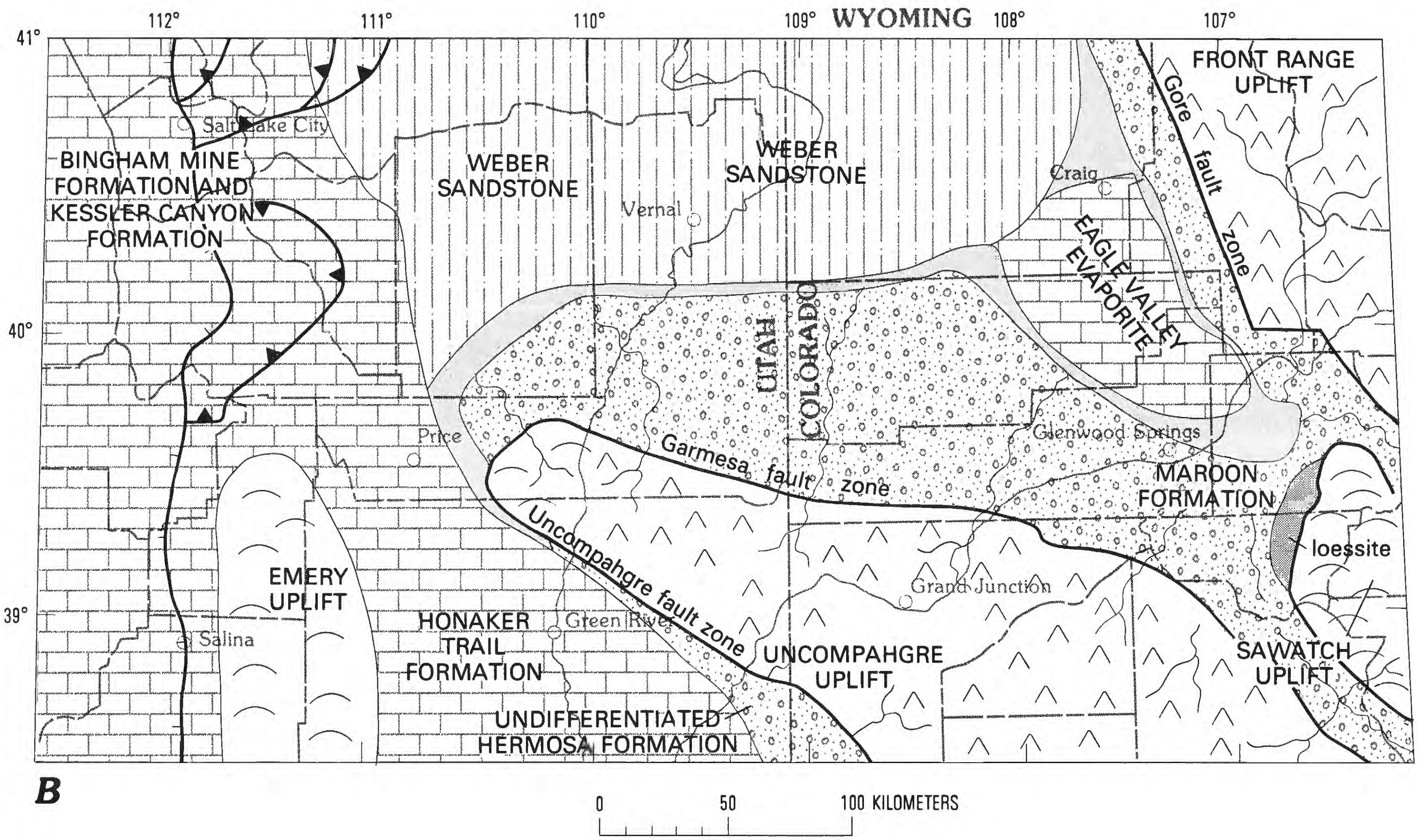
The major uplifts and basins of the Uinta-Piceance region continued to be active during the Missourian and Virgilian (fig. 8). Evaporite sedimentation essentially ended in the Paradox and Eagle basins, probably in response (1) a general rise in sea level (Ross and Ross, 1987, 1988), leading to less restricted circulation, (2) decreasing rates of subsidence in evaporite basins (figs. 10, 11), leading to increased progradation of clastic sediments derived from basin-margin uplifts, and (3) changes in tectonic geomorphology, leading to less topographic restriction of basins. Ross and Ross (1987) suggested there were 17 major eustatic fluctuations during the Missourian and Virgilian, with magnitudes ranging from about 60 to 180 m (fig. 4). Alternatively, Heckel (1986) recognized evidence for 36 major eustatic fluctuations in the North American midcontinent during the Missourian and the lower part of the Virgilian.

Missourian-Virgilian rocks are not preserved on the Emery uplift, and Brill (1963) noted no Missourian and only a thin (<100 m) section of Virgilian limestone on the northern Callville shelf. Brill (1963) suggested that these areas were characterized by minor uplift during most of this interval, whereas Welsh and Bissell (1979) suggested that Missourian and Virgilian strata may have been deposited and then subsequently removed during a latest Pennsylvanian erosional event. Neither hypothesis suggests more than minor subaerial or subaqueous relief in these areas, perhaps close to the inferred magnitude of eustatic fluctuations. Accordingly, our maps show emergence during maximum regressions and partial submergence and carbonate deposition during maximum transgressions. Periodic emergence and erosion of the Callville shelf is consistent with the presence of clasts in correlative rocks in the adjacent Oquirrh basin that include fossils slightly older than their host strata (Jordan and Douglas, 1980).

Relatively rapid subsidence continued in the southeastern Oquirrh basin, where Missourian and Virgilian strata as thick as 2,200 m include the Bingham Mine Formation and the lower part of the Kessler Canyon Formation (fig. 3) (Tooker and Roberts, 1970). In contrast, strata of inferred Missourian and Virgilian age on the east flank of the Oquirrh basin on the Wyoming shelf (the middle part of the Weber Sandstone) are at most a few hundred meters thick (Bissell and Childs, 1958; Welsh and Bissell, 1979).

Both the Bingham Mine and Kessler Canyon Formations consist of sandstone and lesser amounts of interbedded limestone. Sandstone is typically massive or crossbedded and commonly bioturbated (Tooker and Roberts, 1970; Jordan and Douglas, 1980; Jordan, 1981). The *Cruziana* trace-fossil community in sandstones in the Oquirrh Mountain area (Jordan and Douglas, 1980, fig. 8; Jordan, 1981) (fig. 1) suggests a shallow-marine environment subject to wave energy. In the same area, ripple-laminated and crossbedded sandstones were interpreted as shallow-marine tractive-current deposits


WYOMING
FRONT RANGE UPLIFT
Gore fault zone
loessite
SAWATCH UPLIFT
EAGLE VALLEY EVAPORITE
MAROON FORMATION
MAROON FORMATION
Craig
Glenwood Springs
Grand Junction
UNCOMPAHGRE UPLIFT
COLORADO
UTAH
Garmesa fault zone
Uncompahgre fault zone
WEBER SANDSTONE
WEBER SANDSTONE
Vernal
Green River
Green River
UNDIFFERENTIATED HERMOSA FORMATION
HONAKER TRAIL FORMATION
Price
EMERY UPLIFT
Salina
Salt Lake City
BINGHAM MINE FORMATION AND KESSLER CANYON FORMATION
41°
40°
39°
112°
111°
110°
109°
108°
107°
0
50
100 KILOMETERS
A



B

Figure 8 (above and facing page). Maps showing the Missourian to Virgilian paleogeography of the Uinta-Piceance basin region during maximum regression (*A*) and transgression (*B*). See explanation on page CC14.

and massive sandstones as the deposits of density-modified grain flows in deeper water (Jordan, 1981). On our maps (fig. 8), we show shallow-water sandstones as regressive deposits and limestones as transgressive deposits. The transition from deposition of limestone to deposition of deep-water clastic rocks during transgressions (see fig. 9) occurred some time during the Missourian and Virgilian.

The inferred Missourian and Virgilian part of the Weber Sandstone on the Wyoming shelf crops out along the south flank of the Uinta Mountains. Western exposures (west of the Duchesne River area, fig. 1) consist of thick-bedded, rarely crossbedded, calcareous sandstone and minor limestone and dolomitic limestone (Huddle and McCann, 1947; Bissell and Childs, 1958). The sandstones are interpreted as mainly shallow marine in origin but may also in part reflect redeposition of eolian dune sediments deposited during the largest regressions. Accordingly, we show the contact between eolian and shallow-marine facies slightly south of the Duchesne River area. Limestones are the inferred maximum transgression deposits in the western Uinta Mountains area.

East of the Duchesne River area, the inferred Missourian and Virgilian part of the Weber Sandstone is dominated by thick units of massive and crossbedded sandstone, and limestone is uncommon (Huddle and others, 1951; Kinney, 1955; Bissell and Childs, 1958; Bissell, 1964; Fryberger, 1979). The thick sets of crossbedded strata are eolian in origin (Fryberger, 1979) and represent regressive deposition. We infer that the Weber eolian dune field was transgressed during maximum highstands and that eolian sands were redeposited by mass flows to form massive and (or) deformed beds (see Doe and Dott, 1980; Eschner and Kocurek, 1986). This hypothesis is strengthened by the presence of marine fossils in similar sandy facies in the upper (Wolfcampian) part of the Weber Sandstone (Bissell and Childs, 1958; Bissell, 1964). A few transgressions may have resulted in deposition of limestone; however, Driese (1985) argued that Weber limestones in the eastern Uinta Mountains area are nonmarine and formed in interdune ponds. To the south, Weber Sandstone strata interfinger with fluvial (transgressive) and eolian sand-sheet (regressive) deposits of the Maroon Formation that were derived from the northern flank of the Uncompahgre uplift. This facies change is well documented in the Rangely area of northwestern Colorado (for example, Fryberger, 1979).

The Missourian and Virgilian of the northern Paradox basin consists mainly of mixed clastic and carbonate rocks of the Honaker Trail Formation (the upper formation of the Hermosa Group) of Wengerd and Matheny (1958), and the lower part of the "Elephant Canyon Formation" of Baars (1962, 1987, 1991). These strata were considered by Loope and others (1990) as the lower part of the upper member of the Hermosa Formation. Missourian and Virgilian strata in this area may be as thick as about 600 m. Coarse alluvial clastic rocks of the upper part of the Hermosa Formation crop out on the northeastern margin of the basin. Limestone is the most common lithology in the Honaker Trail Formation (for example, Melton, 1972) and is inferred to represent transgressive deposition. Honaker Trail clastic rocks in the northern Paradox basin consist of varicolored siltstone and shale and uncommon sandstone (for example, boreholes 10, 12, 17, fig. 1), which we infer were deposited during regressions in sabkha to shallow-marine environments. Loope (1984) described eolian sandstones from outcrops of the Honaker Trail Formation (he assigned these sandstones to the upper part of the Hermosa Formation) in the Moab area (fig. 1) in the southeastern part of the northern Paradox basin. Thick sandstone beds of possible eolian origin are common in nearby boreholes but pinch out to the north and west. These eolianites are also inferred regressive deposits and may be more abundant in the main part of the Paradox basin to the south.

Relief in the structural zone on the northwestern flank of the Uncompahgre uplift (Potter and others, 1991) is inferred to have been less than in the Desmoinesian, and Missourian and Virgilian strata are inferred to have onlapped the northwest-plunging uplift (for example, borehole 29, fig. 1). This decreased relief probably led to improved circulation through the area of the Oquirrh sag (the eastern Uinta basin) (Szabo and Wengerd, 1975). Several depositional facies converge in and north of the Oquirrh sag, an area for which there is essentially no late Paleozoic outcrop or subsurface control.

In the Eagle basin, Missourian and Virgilian strata include part of the nonmarine Maroon Formation and, in the center of the basin, the uppermost, nonevaporitic part of the Eagle Valley Evaporite (Johnson, 1987a; Johnson and others, 1988). A Missourian and Virgilian age for the upper Eagle Valley Evaporite (not previously inferred) is based on the following. Johnson (1987a) measured a 987-m-thick stratigraphic section of the upper Eagle Valley Evaporite in the Eagle Valley (fig. 1) in the central part of Eagle basin. This section begins at the top of the highest thick (60 m) gypsum bed in the Eagle Valley Evaporite, lacks unconformities, and is a continuation of a stratigraphic section measured by Schenk (1989). Schenk traced the upper Desmoinesian Jacque Mountain Limestone Member, which forms the uppermost bed in the Minturn Formation (fig. 3), from the Minturn area (fig. 1) on the east flank of the Eagle basin into the middle of the Eagle basin, where it immediately underlies the highest thick gypsum. In the Minturn area, Tweto and Lovering (1977) estimated a Maroon Formation thickness of 1,147 m, using the top of the Jacque Mountain Limestone Member as its base. This Maroon Formation thickness is similar to that of the post–Jacque Mountain section of the upper Eagle Valley Evaporite and the Maroon Formation along the Eagle River (1,047 m), and the two sections are considered correlative. The stratigraphic cross section of Johnson and others (1988, fig. 3) shows this relationship and illustrates how the upper Eagle Valley Evaporite in the central part of Eagle basin interfingers with the lower part of the Maroon Formation on both basin margins. Although lacking diagnostic fossils, the Maroon Formation is considered late Desmoinesian to mid-Wolfcampian in age based on regional considerations. If relatively constant sediment accumulation rates are assumed for the Maroon throughout this period, then the upper Eagle Valley Evaporite (which also lacks diagnostic fossils and, as discussed above, interfingers with the lower Maroon) would have a late Desmoinesian to Virgilian age.

The Eagle basin became progressively more nonmarine with time throughout the late Paleozoic, a pattern that probably reflects decreasing rates of subsidence (figs. 10, 11) and associated increased progradation of clastic rocks from the basin margins. Facies and paleocurrent patterns indicate that the axis of Eagle basin was strongly skewed to the northeast during Maroon time (Johnson and others, 1988, fig. 7), constituting a reversal in basin polarity from the Morrowan and Atokan. The Uncompahgre uplift was the main source of clastic detritus in the Maroon (Johnson, 1987b). The Sawatch uplift influenced depositional patterns (Johnson, 1987a, 1989a; Johnson and others, 1988) but probably had low relief.

The Maroon Formation has a maximum thickness of about 1,000 m in the main part of Eagle basin (Johnson and others, 1988) and consists of interbedded fluvial and eolian deposits. Near the basin margins, Maroon fluvial deposits (Johnson, 1987b) consist of channel architectural elements (Miall, 1985) of conglomeratic sandstone. These proximal channel deposits grade down the paleoslope into laminated sand elements of very fine to fine grained sandstone. The facies changes reflect decreases in flow strength, depth, and discharge, characteristics of "terminal fan" fluvial systems in arid basins (Friend, 1978).

Maroon Formation eolianites are mainly sand-sheet deposits (Johnson, 1987a; Johnson and others, 1988). These sand-sheet deposits typically comprise about 30 percent of Maroon sections and appear to cyclically alternate with fluvial intervals (fig. 5). Some fluvial deposits (shown schematically in fig. 8) were also deposited lateral to the eolian sand sheets, in wadi streams. Silt eroded from the sand-sheet deposits in the Maroon Formation was transported to the downwind basin margin, the northwest flank of the Sawatch uplift (Ruedi Reservoir area, fig. 1), where it was redeposited as a thick section of loess (Johnson, 1989a).

Johnson (1987a) and Johnson and others (1988) suggested that the inferred nonmarine cyclicity in the Maroon Formation was controlled by cyclic climatic changes synchronous with global glacio-eustatic changes. In this interpretation, fluvial deposition dominated in the Maroon during wetter climatic periods (correlating with eustatic highstands), and eolian sand-sheet deposition dominated during drier climatic periods (eustatic lowstands).

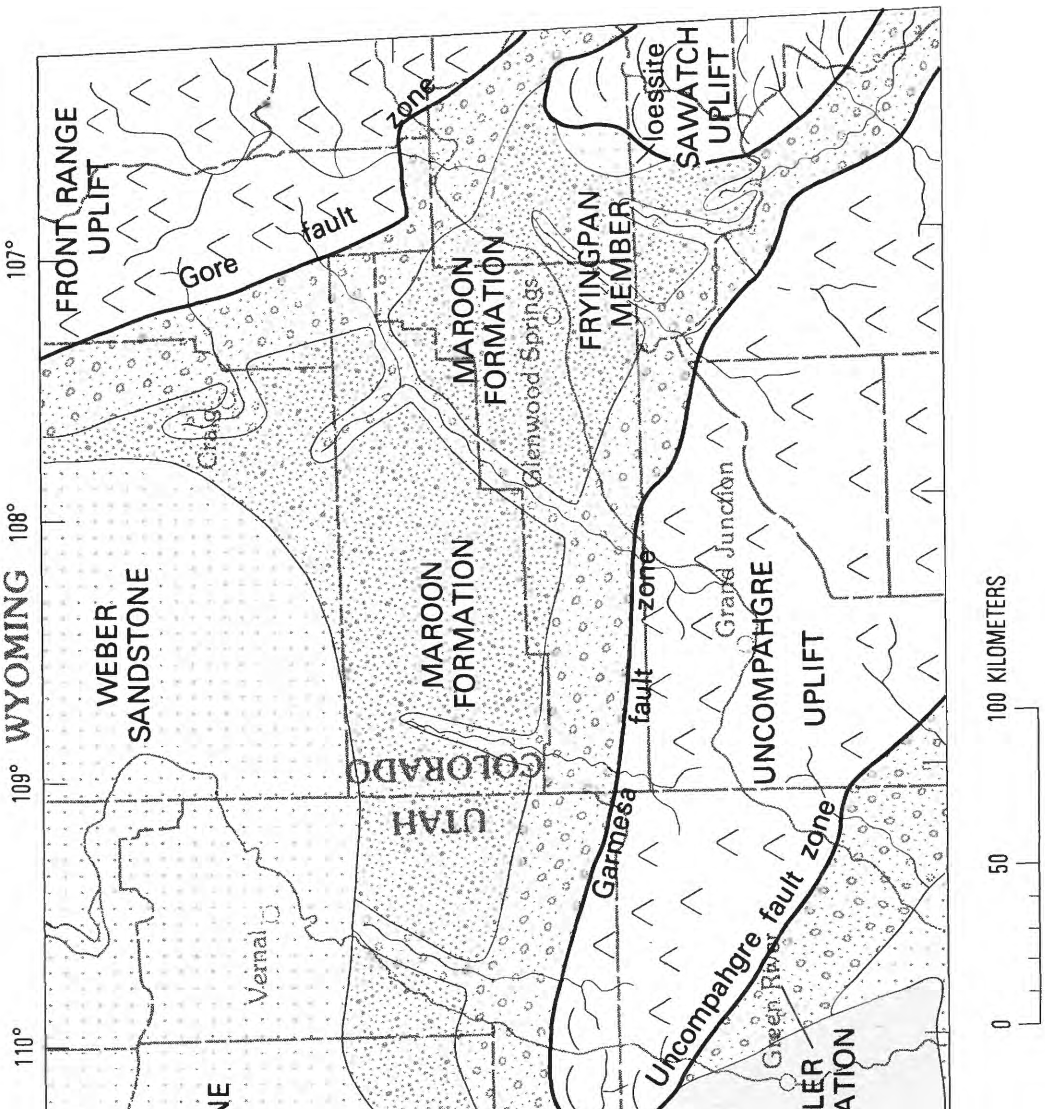
The upper, nonevaporitic part of the Eagle Valley Evaporite, which crops out along the Missourian-Virgilian axis of the Eagle basin, consists of several gradational facies (Johnson, 1987a; Johnson and others, 1988): (1) plane-bedded or low-angle-bedded, varicolored, fine- to coarse-grained sandstone; (2) ripple-laminated siltstone to fine-grained sandstone; (3) massive, extensively bioturbated sandstone and siltstone; (4) grayish-black, thinly laminated, calcareous siltstone and mudstone; and (5) grayish-black micritic limestone and lime mudstone. These facies represent a continuum from sabkha and shallow-water (regressive) to offshore marine (transgressive) deposition. As in the late Atokan and Desmoinesian, the Eagle basin was probably completely cut off from normal marine waters during regressions, when the central part of the basin may have resembled a playa lake. During transgressions, marine waters entered the basin from the north. Marine limestone of inferred Missourian and Virgilian age (based on stratigraphic position) is present at Ripple Creek, Miller Creek, and Sylvan (Johnson, 1987a) and in boreholes 40, 41, 45, 48, and 49 (fig. 1). The contact between these limestones and sandy marine environments (the transgressed Weber sand sea) in northwestern Colorado is between boreholes 40 and 50.

In summary, the Missourian and Virgilian in the Uinta-Piceance region was characterized by decreasing tectonic activity, reflected in lower rates of subsidence (figs. 10, 11). In the Eagle basin, fluvial and eolian deposition dominated while evaporite deposition ended. Evaporite deposition similarly ended in the Paradox basin and was replaced by deposition of sabkha and (or) shallow-marine deposits during regressions. Progradation of sands across the Wyoming shelf continued during regressions. Transgressive deposition of limestone was limited to the western part of the Uinta-Piceance region and a small area in the eastern Eagle basin.

Early Early Permian—Early Wolfcampian

The major basins of the Uinta-Piceance region (fig. 9) continued to subside during the early Wolfcampian (figs. 10, 11). As the Emery uplift subsided, however, the Paradox basin became connected to the Callville shelf and its history as a narrow trough ended. Subsidence exceeded sediment supply in the southeastern Oquirrh basin, so that it was both a depositional and topographic basin. Depositional patterns on the Wyoming shelf and in the Eagle basin remained similar. Ross and Ross (1987, 1988) recognize 11–15 major eustatic changes in the Wolfcampian, with inferred magnitudes of as much as 170 m (fig. 4).

The Pakoon Limestone (50–70 m thick) represents early Wolfcampian deposition on the Callville shelf; the Queantoweap Sandstone (50–90 m thick) represents late Wolfcampian deposition (Brill, 1963; Welsh, 1976). Few lithologic descriptions are available for the Pakoon in the Uinta-Piceance region. It apparently consists of dolomitic limestone and minor sandstone (borehole 1 and Cove Fort area outcrops, fig. 1), inferred to represent transgressive and



WYOMING
UTAH
COLORADO
FRONT RANGE UPLIFT
Gore fault zone
SAWATCH UPLIFT
loessite
FRYINGPAN MEMBER
MAROON FORMATION
MAROON FORMATION
WEBER SANDSTONE
WEBER SANDSTONE
UNCOMPAHGRE UPLIFT
Garmesa fault zone
Uncompahgre fault zone
CUTLER FORMATION
ELEPHANT CANYON FORMATION
OQUIRRH GROUP
PAKOON DOLOMITE
Craig
Glenwood Springs
Grand Junction
Vernal
Green River
Price
Salt Lake City
Salina
107°
108°
109°
110°
111°
112°
41°
40°
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50
100 KILOMETERS
A

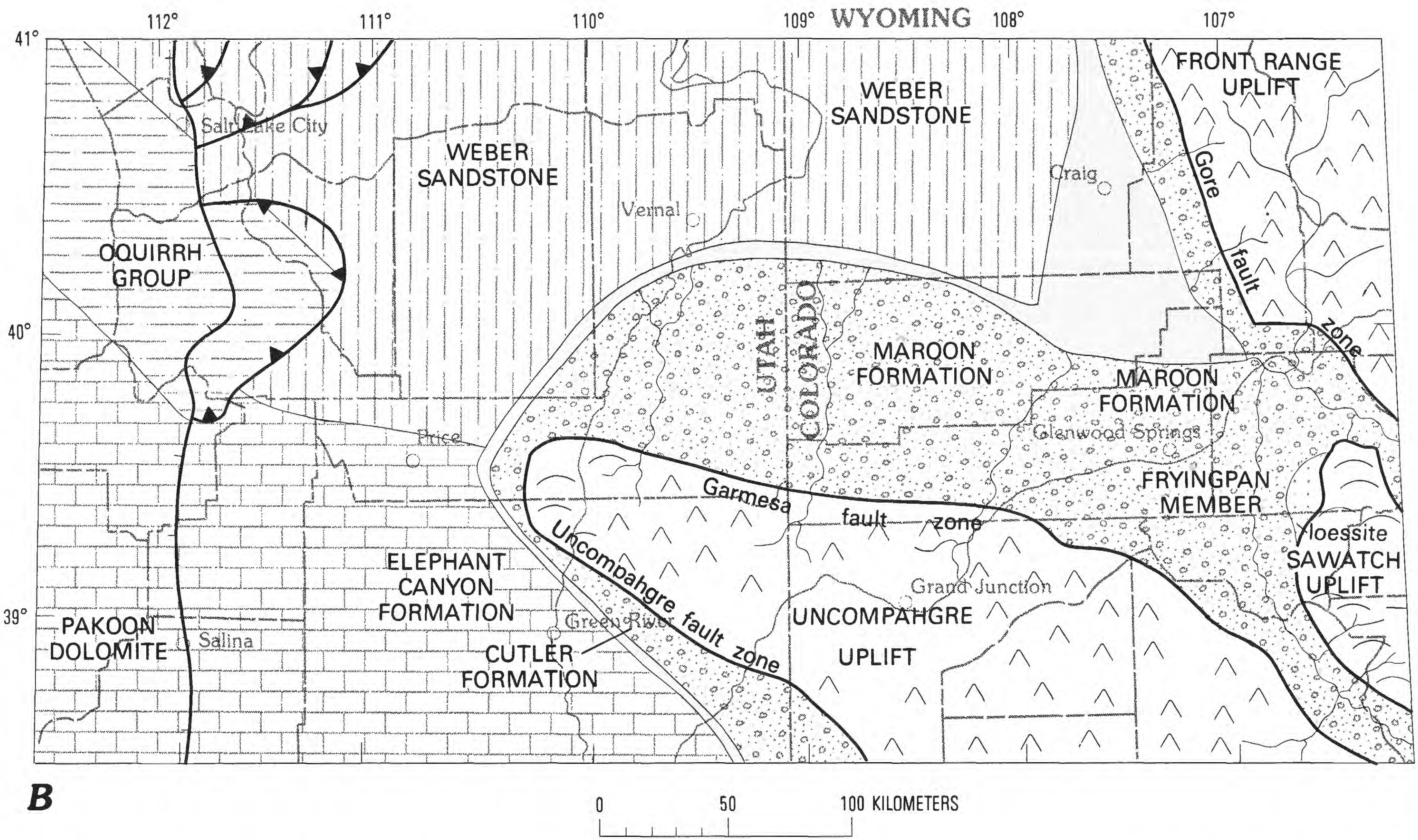


Figure 9 (above and facing page). Maps showing the early Wolfcampian paleogeography of the Uinta-Piceance basin region during maximum regression (*A*) and transgression (*B*). See explanation on page CC14.

regressive deposition, respectively. J.E. Welsh (oral commun. to M.A. Chan, 1989) suggested that features of the Pakoon carbonate rocks indicate a somewhat restricted, shallow-marine environment. George (1985) reported that Pakoon sandstones are crossbedded and that they resemble sandstones in the unconformably underlying Callville Limestone. These sandstones could be shallow marine (as shown in fig. 9) or possibly eolian in origin. Because Wolfcampian subsidence rates on the Callville shelf were apparently very low (suggested by the small thickness of preserved section), it is also possible that the Callville shelf was an emergent, erosional area during larger regressions. There are apparently no transitional facies between Pakoon and Queantoweap strata in the Cove Fort area and the thick section of Oquirrh basin Wolfcampian strata to the north in the Route 148 southwest area (Brill, 1963; Morris, 1977) (fig. 1).

The Pakoon Limestone apparently grades eastward into the upper part of the Elephant Canyon Formation of Baars (1962, 1987, 1991) in the northern Paradox basin; these latter strata were considered the Rico Formation by Loope (1984) and the upper part of the upper member of the Hermosa Formation and "lower Cutler beds" by Loope and others (1990). Although debate concerning the stratigraphic nomenclature continues (for example, Baars, 1991), the name Elephant Canyon Formation is used here for convenience and because of its prevalence in the literature. In the main part of the northern Paradox basin, the Elephant Canyon Formation is as thick as about 460 m and consists of interbedded carbonate rocks, fine-grained clastic rocks, and minor anhydrite (for example, boreholes 8, 14, 16, 17, 24, 26, fig. 1) (Baars, 1962, 1987; Terrell, 1972; Loope, 1984). Limestones are commonly fossiliferous and are interpreted as shelf sediments deposited during transgressions. Interbedded clastic rocks are interpreted as sabkha, playa, and shallow-marine regressive deposits.

Sandstones of inferred eolian origin are also present at a few locations in the upper part of the Elephant Canyon Formation in the southern part of the northern Paradox basin (for example, boreholes 4 and 6, fig. 1) and in a narrow belt near the contact with the Cutler Formation on the east flank of the basin (Terrell, 1972; Loope, 1984; Campbell, 1987). Chan and Langford (1987) described cycles (about 6–40 m thick) consisting of marine and erg-margin deposits in the upper part of the Elephant Canyon Formation in the central to northern Paradox basin. Transgressive deposits consist of normal-marine fossiliferous shelf limestone and tidal sandstone and are as thick as a few tens of meters. Regressive deposits consist of fluvial and eolian sandstone. The relatively sharp lower contacts of limestone beds indicate rapid eustatic rises and inundation of erg margins. The marine incursions are thought to have entered the Paradox basin from the west (Baars, 1979; Campbell, 1987). Closely spaced (every 4–15 m) supersurfaces in the eolian and fluvial units may have formed during wet climatic intervals by river flooding (Langford and Chan, 1988, 1990). Based on the abundance of the supersurfaces, Chan and Langford (1987) suggested that climatic fluctuations of shorter periodicity than late Paleozoic eustatic fluctuations may have influenced sedimentation in this area.

The Cutler Formation comprises coarse alluvial conglomerate and arkosic sandstone derived from the Uncompahgre uplift. As in the Missourian and Virgilian, narrowing of the width of the band of coarse clastics derived from this highland to the northwest suggests diminished relief on the Uncompahgre in that direction. Onlap of the northwestern flank of the Uncompahgre uplift apparently continued in the Wolfcampian.

In the area of the Emery uplift, transitional strata between the Pakoon Limestone to the west and the upper part of the Elephant Canyon Formation to the east consist of dolomite and limestone, thick to thin-bedded sandstone, varicolored siltstone and shale, and minor anhydrite (boreholes 2, 3, 5, 19–23, 25, fig. 1). The proportion of sandstone is variable but probably is significantly greater than in the upper part of the Elephant Canyon Formation. Most likely the sands were not transported westward across the Paradox basin from the Uncompahgre uplift but instead had northern sources in sand-rich Weber environments of the Wyoming shelf or southern sources in the eolian sands of the Cedar Mesa Sandstone of Baars (1962). The sedimentology of these Elephant Canyon sands is unknown, but we infer that they were deposited by eolian processes during larger regressions and graded westward into shallow-marine environments. The Cedar Mesa Sandstone interfingers with the upper part of the Elephant Canyon Formation to the south in the main part of the Paradox basin and overlies the upper Elephant Canyon Formation in the northern Paradox basin.

Lower Wolfcampian strata in the southeastern Oquirrh basin include the Curry Peak and Freeman Peak Formations and undifferentiated strata of the Oquirrh Group (Welsh and James, 1961; Jordan and Douglas, 1980). These Wolfcampian strata are as thick as 1,500 m thick and are unconformably overlain (in sequence) by the Diamond Creek Sandstone of Wolfcampian age and the Park City or Phosphoria Formations of late Early (Leonardian to Wordian) Permian age. Lower Wolfcampian rocks are mainly massive, crossbedded, graded, and ripple-laminated sandstone and minor interbeds of conglomerate and limestone (Larson, 1979; Larson and Clark, 1979; Jordan and Douglas, 1980; Jordan, 1981). Conglomerate crops out in a northwest-trending band along the southwest margin of the basin and contains limestone clasts older than host strata. Chamberlain and Clark (1973), Larson (1979), Jordan (1979, 1981), and Jordan and Douglas (1980) interpreted the sandstones and conglomerates as the deposits of sediment gravity flows, including turbidity currents, density-modified grain flows, and debris flows. Trace-fossil communities in the sandstones suggest that deposition occurred below wave base at bathyal depths (Chamberlain and Clark, 1973; Larson, 1979; Larson and Clark, 1979; Jordan and Douglas, 1980; Jordan, 1981).

Jordan and Douglas (1980) suggested a slope or basin-floor depositional environment at depths of a few hundred meters or more. These depths exceed the magnitude of inferred sea-level changes (Ross and Ross, 1987) and thus provide a rationale for similar depositional processes and environments during both eustatic highstands and lowstands. Lowstand deposits are probably more conglomeratic, containing clasts derived from periodically exposed platforms and carried basinward through erosional channels (see Larson and Clark, 1979). On figure 9, the location of the transition from shallow- to deep-water environments on the northeastern basin flank is based on the maps of Jordan and Douglas (1980) west of the Wasatch fault and on the facies descriptions of Larson (1979) and Larson and Clark (1979) in the area of the Charleston-Nebo thrust sheet. On the southwestern flank of the basin, this transition must lie northeast of the Lower Permian limestones of the Oquirrh Group mapped by Morris (1977) in the Route 148 southwest area (fig. 1).

The upper part of the Weber Sandstone consists almost entirely of sandstone and comprises the early Wolfcampian of the Wyoming shelf depositional province (Bissell and Childs, 1958). It is unconformably overlain by the Park City Formation. Sandstones of the Weber are predominantly crossbedded in the eastern Uinta Mountains (east of the Duchesne River area, fig. 1) and predominantly massive in the western Uinta Mountains. These strata were probably deposited in eolian (maximum regressions) and shallow-marine (maximum transgressions) environments, and the relative proportions of the two bedding types reflect the dominant depositional process. Shallow-marine conditions probably dominated in the western Uinta Mountains; eolian dunes probably formed only during a few of the larger regressions and were rapidly reworked in subsequent transgressions. Similarly, the large proportion of crossbedded strata in the eastern Uinta Mountains suggests that eolian conditions dominated and that only the largest transgressions submerged the Weber dune field. The presence of marine fossils (Bissell and Childs, 1958), massive beds, contorted stratification, and extensive burrowing (Fryberger, 1979) all support intermittent marine conditions for the Weber in the eastern Uintas. The Weber Sandstone interfingers to the south and southeast with fluvial and eolian sand-sheet deposits of the Maroon Formation (boreholes 31–35, fig. 1), which were derived from the northern flank of the ancestral Uncompahgre uplift.

The upper part of the Maroon Formation comprises the early Wolfcampian of the Eagle basin. These strata are unconformably overlain by the Upper Permian to Lower Triassic State Bridge Formation. Similar to the Missourian and Virgilian lower part of the Maroon, these strata consist of mixed fluvial and eolian sand-sheet deposits in most of the basin. The basin continued to be asymmetric, with its depositional axis skewed strongly to the east. During transgressions, fluvial systems draining the Uncompahgre and Front Range uplifts flowed across the basin, converged along the depositional axis, and flowed north to exit the basin (Johnson, 1987a, b; Johnson and others, 1988). Johnson (1987a) described rare marine beds from the upper Maroon in the Sylvan, Miller Creek, and Ripple Creek areas (fig. 1), and Hallgarth (1959) showed dolomitic and gypsiferous mudstones from the uppermost Maroon in boreholes 41, 45, and 47 (fig. 1). These strata were probably deposited during the largest transgressions in a sabkha or marginal-marine setting.

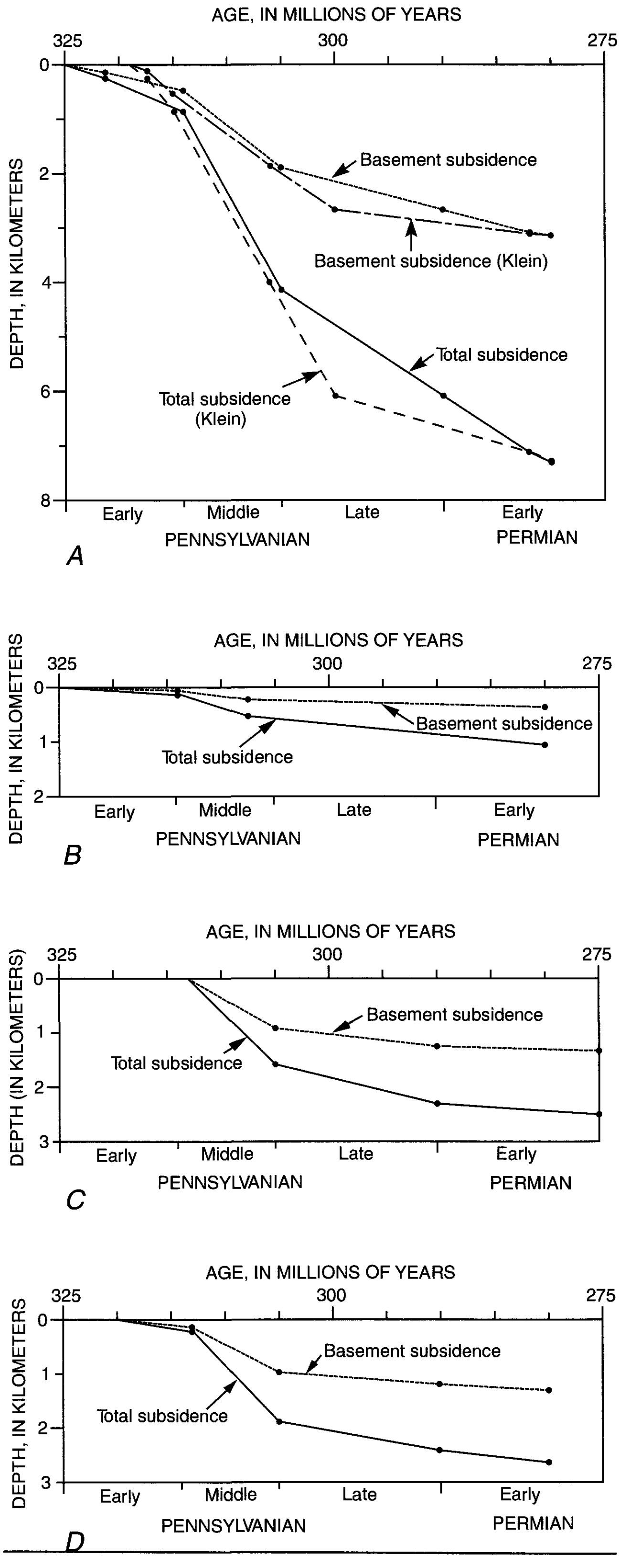
As in the Missourian and Virgilian, eolian sand-sheet deposits were dominant in the Eagle basin during regressions and merged to the north with the Weber dune deposits. Loess continued to be deposited in the Ruedi Reservoir area (fig. 1) on the downwind basin margin adjacent to the low-relief Sawatch uplift. Near the end of Maroon deposition, however, deposition of eolian dune sediments of the Fryingpan Member of the Maroon Formation over the loessite formed a basin-margin dune field (Johnson, 1989b). The termination of loess deposition and the initiation of dune deposition were probably forced by a cessation or slowing of subsidence in the Eagle basin, the source area for the basin-margin eolianites. Rather than being rapidly buried beneath the Maroon alluvial-eolian plain, sands in the source area were exposed at the surface and made susceptible to eolian erosion and transport for longer time intervals. Consistent with this interpretation, thick beds of coarse to granular sand interpreted as deflation lags are present at the top of the Maroon Formation in many places (Johnson and others, 1988).

In summary, depositional patterns initiated in the Missourian and Virgilian in the Eagle basin and on the Wyoming shelf continued into the early Wolfcampian. The Emery uplift became fully or mostly submerged and the Paradox basin ceased to be a discrete geomorphic element. Deep-water clastic deposition began in the Oquirrh basin (or continued from the latter part of the Missourian and Virgilian). Transgressive limestone deposition was limited to the southeastern part of the Uinta-Piceance basin region.

DISCUSSION

The geohistory diagrams (figs. 10, 11) show that the Oquirrh basin experienced the greatest rates and magnitude of subsidence, the Eagle and northern Paradox basins experienced intermediate and similar rates, and the Wyoming shelf experienced the lowest rates and magnitude of subsidence. All four sedimentary provinces experienced their lowest rates of subsidence in the Early Pennsylvanian, their highest rates in the Middle Pennsylvanian, and intermediate rates in the Late Pennsylvanian and Early Permian. The decrease in subsidence rates from the Middle to Late Pennsylvanian in the Oquirrh basin is minor in contrast to decreases in the other three depositional provinces.

Subsidence in the northern Paradox basin (fig. 10*C*) is inferred to have been mainly controlled by flexural loading



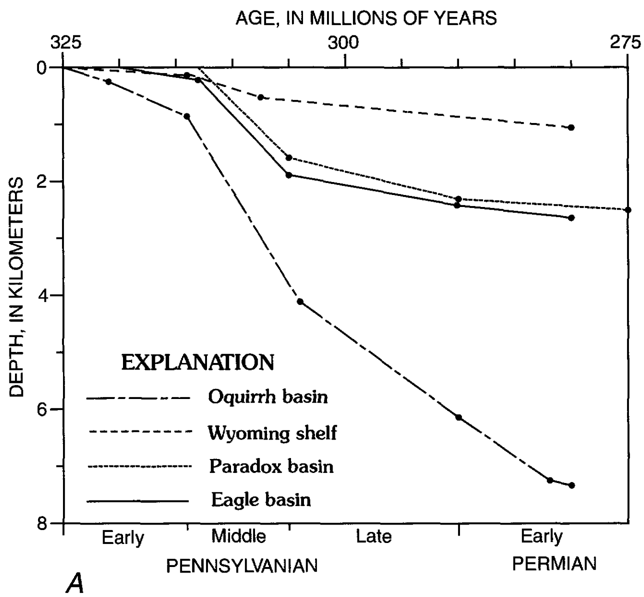
driven by westward overthrusting of the Uncompahgre uplift. As stated earlier, the basin axis was skewed strongly to the east adjacent to the Uncompahgre uplift (Mallory,

Figure 10 (facing column). Geohistory diagrams for four sedimentary provinces in the Uinta-Piceance basin region. *A*, Oquirrh basin; *B*, Wyoming shelf; *C*, northern Paradox basin; *D*, Eagle basin. The origin of each curve at the beginning of the Pennsylvanian is assumed to be at sea level because the Late Mississippian in each province was characterized either by shallow-marine carbonate deposition (*A*, *B*) or by minimal emergence (*C*, *D*). Solid lines represent total subsidence. Dotted lines represent subsidence of the Late Mississippian "basement" corrected for the incremental load induced by the weight of sediment through time and, thus, the subsidence due to tectonics. Solid circles represent the time-thickness data points used in the analysis. Time scale used is from Haq and Van Eysinga (1987). Diagram *A* also shows the same data plotted using the time scale proposed by Klein (1990). Corrections for compaction are based on lithology and follow the exponential porosity function presented by Sclater and Christie (1980). No corrections were made for bathymetry or eustatic fluctuations. Because of uncertainties involving the ages of units and their compaction and diagenetic histories, the plots should be regarded as approximations. Sources of data: *A*, south-central Oquirrh Mountains (Tooker and Roberts, 1970); *B*, Split Mountain anticline (Kinney, 1955; Thomas and others, 1945); *C*, northern Paradox basin (Jay Roberts no. 1 Whitecloud, sec. 34, T. 22 S., R. 17 E.); *D*, White River uplift (Johnson and others, 1988).

1972, 1975), consistent with this inference. Subsidence in the eastern Eagle basin (fig. 10*D*) closely resembles that in the northern Paradox basin, supporting the contention that the Gore fault (for which the sense of late Paleozoic offset is uncertain) on the eastern margin of Eagle basin was also a reverse fault. Alternatively, two different yet synchronous tectonic processes (flexural loading in the northern Paradox basin, oblique or normal rifting in the Eagle basin) would need to be invoked to produce their almost identical subsidence histories.

Evaporite deposition in both the northern Paradox basin and the Eagle basin probably was controlled by subsidence rate, drainage restriction, arid climate, and fluctuating eustasy. Evaporite deposition in each basin occurred only during the Middle Pennsylvanian, the time of most rapid subsidence. Rapid subsidence commonly results in trapping of clastic sediment adjacent to basin margins (Blair and Bilodeau, 1988; Heller and others, 1988), causing basin interiors to become isolated from supplies of clastic sediment (or sediment starved), a partial requirement for evaporite deposition. Drainage restriction took different forms in the two basins, with an inferred tectonic barrier at the northern end of the Paradox basin and a combined tectonic (slowly subsiding) and sedimentary (the positive relief on the Morgan Formation dune field) barrier at the northern end of the Eagle basin (fig. 7*A*). Regular eustatic fluctuations led to submergence of these barriers and replenishment of the water column. Re-emergence of barriers during eustatic regressions led to isolation of water bodies and climate-induced evaporite deposition. All of the above conditions were required for evaporite deposition; evaporite deposition ceased in the Late Pennsylvanian when subsidence rates diminished (figs. 10*C*, D, 11), and clastic sediments began to prograde into the basin interiors.

The slowly subsiding Wyoming shelf (fig. 10B) served as a bypass zone across which sediment was transported southwestward to the rapidly subsiding Oquirrh basin. The much larger amount of total subsidence in the Oquirrh basin (6–7 times greater than on the Wyoming shelf) (fig. 11) suggests that significant structures (faults or ramps) were present between the two depositional provinces. This inferred zone of structural weakness is not presently recognizable; it was presumably reactivated during the Cretaceous Sevier and early Tertiary Laramide orogeny or has been buried by younger sediments. Despite the contrasts in subsidence history, shallow-marine deposition in both the Wyoming shelf and the Oquirrh basin through most of the Pennsylvanian indicates that sediment supply kept up with subsidence in the Oquirrh basin. Deepening of the Oquirrh basin in the latest Pennsylvanian and the Early Permian (fig. 9) was apparently not accompanied by a significant increase in subsidence rates (fig. 10A). Diminishing sediment supply may therefore have led to the formation of deep-water depositional environments.



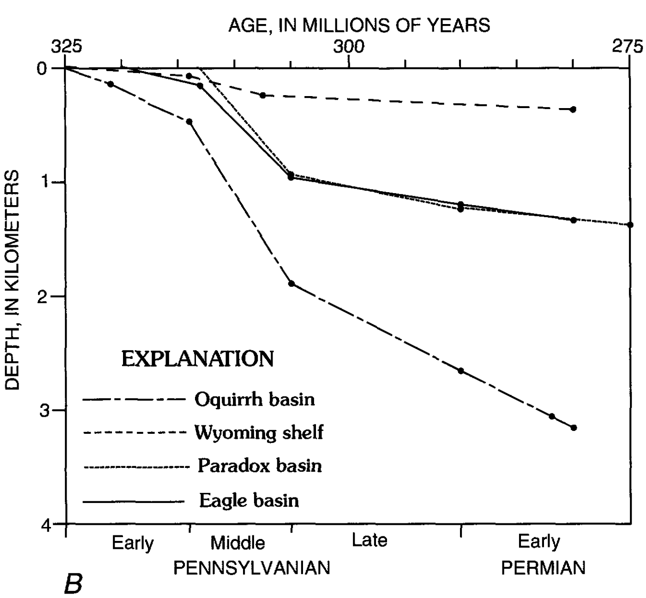


Figure 11. Geohistory diagrams comparing total subsidence (A) and basement subsidence (B) in four sedimentary provinces in the Uinta-Piceance basin region. See figure 10 for details concerning construction of the diagrams.

The Oquirrh basin is distinguished from the three other depositional provinces plotted on figure 11 on the basis of its higher rates and magnitude of subsidence and its continuing high subsidence rates through the Late Pennsylvanian and the Early Permian. All of the Uinta-Piceance basin region was located east of the I_{Sr}=0.706 isopleth for Mesozoic and Cenozoic igneous rocks (Kistler and Peterman, 1978); thus its basement rocks are similar. The Oquirrh basin, however, does overlie the hinge zone of the Late Proterozoic to mid-Paleozoic miogeocline (Hill, 1976), and weakened crust in this zone might in part be responsible for the anomalously high subsidence in the Oquirrh basin. More importantly, the tectonic controls on subsidence in the Oquirrh basin were probably different from those in the Eagle and Paradox basins.

Both the timing (figs. 10, 11) and geometry of subsidence suggest that late Paleozoic deformation in the Uinta-Piceance region of the ancestral Rocky Mountains resulted from interactions along both the western and southeastern continental margins (fig. 2). As stated earlier, initial subsidence of the Oquirrh basin and the Wyoming shelf as discrete basinal elements began in the Late Mississippian. This subsidence predates the onset of significant deformation in the more proximal parts of the foreland province associated with the Marathon-Ouachita orogenic belt (the collisional mountain belt of the southeastern convergent margin) by 10–20 million years and predates initial subsidence in the Eagle and Paradox basins by 25–30 million years (Kluth, 1986). Assuming that the effects of the continent-continent collision to the southeast propagated inland with time as the collision broadened or even that deformation was initiated synchronously across the entire foreland, the early phases of subsidence in the Oquirrh basin and Wyoming shelf are too old to be the result of the collision and demand an independent driving force. The inferred extensional or transtensional origin of the Oquirrh basin is consistent with the structural style noted to the west by Ketner (1977) and Smith and Miller (1990), and a western driving force is highly likely. In contrast, the inferred contractional or transpressional style of the Paradox basin and possibly the Eagle basin may be more consistent with the foreland deformation associated with a convergent margin.

Can the effects of the western and southeastern driving forces be isolated? Because significant subsidence in the Eagle and Paradox basins did not begin until the Atokan, these basins may never have been significantly affected by the western driving force. The Desmoinesian acceleration of subsidence in the Eagle, Paradox, and Oquirrh basins is roughly correlative and therefore may in each case reflect the growing influence of the southeastern force (the dominant influence in the Eagle and Paradox basins and a supplemental influence in the Oquirrh basin). The Late Pennsylvanian decrease in subsidence in the Eagle and Paradox basins indicates that the influence of the southeastern force diminished; continued rapid Late Pennsylvanian and Early Permian subsidence in the Oquirrh basin supports the continued importance of west-induced extensional deformation in the western Uinta-Piceance region during this interval.

While considering only interactions on the southern continental margin as a driving force for deformation in the ancestral Rocky Mountains, Kluth (1986) noted the high angle between the uplifts and basins of the ancestral Rocky Mountain orogeny and the Marathon-Ouachita orogenic belt. Although Kluth presented a rationale for these anomalous trends in the context of northward-imposed paleostress, the geometry of structures in at least the western ancestral Rocky Mountains is more consistent with paleostress imposed by transtensional deformation along a western continental margin. Regional transtension in the ancestral Rocky Mountain province is supported by the mix of compressional and extensional structural styles. Future work on Pennsylvanian and Permian deformation and subsidence in the western United States (including locating and describing the structures bounding the Oquirrh basin), along with better documentation of modern analogues, will further clarify the nature of intraplate deformation in the ancestral Rocky Mountains.

CONCLUSIONS

Expansion and contraction of late Paleozoic Gondwana ice sheets resulted in repetitive global eustatic and climatic fluctuations. These eustatic and climatic fluctuations, along with tectonic activity and variations in sediment supply, were major controls on the development of transgressive-regressive depositional sequences in the Uinta-Piceance basin region of northwestern Colorado and northeastern Utah. Major changes in depositional patterns in the Uinta-Piceance region associated with the repeated transgressions and regressions are shown on four sets of paleogeographic maps representing key time periods in the late Paleozoic. The contrasts between maps representing maximum transgressions and regressions are substantial and indicate the necessity of this approach for adequately representing late Paleozoic paleogeography.

The Uinta-Piceance basin region includes four major sedimentary provinces: the Eagle basin, the northern Paradox basin, the southern Wyoming shelf, and the southeastern Oquirrh basin. The Oquirrh basin experienced the greatest rates and magnitude of subsidence, the Eagle and northern Paradox basins experienced intermediate and very similar rates, and the Wyoming shelf experienced the lowest rates and magnitude of subsidence. All four sedimentary provinces experienced their lowest rates of subsidence in the Early Pennsylvanian, their highest rates in the Middle Pennsylvanian, and intermediate rates in the Late Pennsylvanian and Early Permian. The timing, magnitude, and geometry of subsidence in the Uinta-Piceance region suggests that regional transtension was driven by the overlapping influences of a more distant, convergent continental margin to the southeast and a more proximal, transform-fault (?) continental margin to the west. The effects of this complex intraplate deformation are reflected in the considerable variations within transgressive-regressive depositional sequences.

Morrowan and lower Atokan strata throughout most of the Uinta-Piceance basin region consist of alternating fine-grained clastic rocks (regressive deposits) and more abundant limestone (transgressive deposits). Significant Morrowan and early Atokan tectonic uplifts include the ancestral Front Range and Sawatch uplifts. Intrabasinal faulting characterized the Eagle basin (Waechter and Johnson, 1986) and locally influenced deltaic and prodeltaic deposition.

The late Atokan to Desmoinesian history of the Uinta-Piceance region was characterized by increased tectonic activity, which is reflected in higher rates of basin subsidence, continued unroofing of the Front Range and Sawatch uplifts, and initial uplift of the Uncompahgre uplift. The combined effects of tectonism and eustasy led to restricted circulation and evaporite deposition in the Eagle and Paradox basins. Regressive deposition was also characterized by significant progradation of eolian sands across the Wyoming shelf en route to the Oquirrh basin. Limestone deposition dominated during transgressions.

The Missourian to Virgilian history of the Uinta-Piceance basin region was characterized by decreasing tectonic activity and rates of subsidence. Progradation of clastic sediments into the central parts of the Eagle and northern Paradox basins resulted in cessation of evaporite deposition. During regressions, fluvial and eolian deposition dominated in the Eagle basin, whereas sabkha and (or) shallow-marine deposition dominated in the northern Paradox basin. The Wyoming shelf continued to serve as a conveyor belt of clastic sediment to the Oquirrh basin, which was characterized by a transition from shallow- to deep-water deposition. Transgressive deposition of limestone was limited to the western part of the Uinta-Piceance basin region and a small area in the eastern Eagle basin.

Depositional patterns initiated in the Missourian and Virgilian in the Eagle basin and on the Wyoming shelf continued into the early Wolfcampian. The Emery uplift was fully or mostly submerged, and the Paradox basin ceased to be a discrete geomorphic element. Deep-water clastic deposition continued in the Oquirrh basin. Transgressive limestone deposition was limited to the southwestern part of the Uinta-Piceance region.

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Appendix—Borehole Geophysical Logs

[Locations shown by number on fig. 1]

1. Shell Oil Company Sunset Canyon (W) No. 1, sec. 21, T. 22 S., R. 4 W., Millard County, Utah.
2. Phillips Petroleum Company United States "E" No. 1, sec. 27, T. 19 S., R. 3 E., Sanpete County, Utah.
3. Skelly Oil Company Emery Unit No. 1, sec. 34, T. 22 S., R. 5 E., Sevier County, Utah (Irwin, 1976).
4. Mountain Fuel Supply South Last Chance Unit No. 1A, sec. 18, T. 26 S., R. 7 E., Emery County, Utah.
5. Pan American Oil Company No. 3 Ferron Unit, sec. 21, T. 20 S., R. 7 E., Emery County, Utah.
6. Pan American Petroleum Corporation No. 9 Nequoia Arch, sec. 25, T. 26 S., R. 13 E., Emery County, Utah.
7. Delhi Oil Company No. 1 Russell, sec. 34, T. 25 S., R. 12 E., Emery County, Utah.
8. Superior Oil Company North Springs Wash No. 31–15, sec. 15, T. 25 S., R. 15 E., Emery County, Utah.
9. Jay Roberts No. 1 Whitecloud, sec. 14, T. 26 S., R. 20 E., Grand County, Utah (Hite and Liming, 1972).
10. Hilliard Oil and Gas Company Klondike No. 1, sec. 24, T. 23 S., R. 18 E., Grand County, Utah.
11. Texaco Oil Company No. 1 Government MacKinnon, sec. 15, T. 23 S., R. 19 E., Grand County, Utah.
12. Superior Oil Company No. 22–34, sec. 34, T. 22 S., R. 17 E., Grand County, Utah (Hite and Liming, 1972).
13. Amax Petroleum Company No. 24–1 Green River Desert, sec. 24, T. 22 S., R. 13 E., Emery County, Utah.
14. Superior Oil Company No. 14–24, sec. 24, T. 21 S., R. 15 E., Emery County, Utah.
15. Gulf Oil Company No. 1 Norris, sec. 8, T. 18 S., R. 16 E., Emery County, Utah.
16. Humble Oil and Refining Company No. 2 Woodside Unit, sec. 30, T. 18 S., R. 14 E., Emery County, Utah.
17. Placid Oil Company No. 1 Marsh Flat, sec. 29, T. 17 S., R. 14 E., Emery County, Utah.
18. El Paso Natural Gas Company No. 1 Pack Saddle, sec. 12, T. 18 S., R. 12 E., Emery County, Utah.
19. Jake Hamon No. 8–1 USA Federal, sec. 8, T. 19 S., R. 9 E., Emery County, Utah.
20. Tiger Oil Company No. 4 1–15 Jack Curtis, sec. 15, T. 18 S., R. 7 E., Emery County, Utah.
21. Phillips Petroleum Company Huntington Anticline No. 1, sec. 15, T. 17 S., R. 8 E., Emery County, Utah.
22. Atlantic Richfield No. 1 Hiawatha, sec. 13, T. 15 S, R. 7 E., Emery County, Utah.
23. Tenneco Oil Company Clear Creek No. 1, sec. 17, T. 14 S., R. 7 E., Carbon County, Utah.
24. Shell Oil Company No. 1 North Springs Unit, sec. 27, T. 15 S., R. 9 E., Carbon County, Utah.
25. Pure Oil Company No. 1–A Washboard, sec. 12, T. 16 S., R. 9 E., Emery County, Utah.
26. Pure Oil Company No. 1 Desert Lake, sec. 1, T. 17 S., R. 10 E., Emery County, Utah.
27. Shell Oil Company No. 1 Miller Creek, sec. 26, T. 15 S., R. 10 E., Carbon County, Utah.
28. Pan American Oil Company Federal Mounds No. 1, sec. 11, T. 16 S., R. 11 E., Emery County, Utah.
29. Shell Oil Company No. 1–A Farnham Dome, sec. 8, T. 15 S., R. 12 E., Carbon County, Utah.
30. Mountain Fuel Supply No. 1 Sunnyside, sec. 17, T. 15 S., R. 13 E., Carbon County, Utah.
31. Chevron Oil Company No. 1 Stone Cabin, sec. 29, T. 12 S., R. 15 E., Carbon County, Utah.
32. Gulf Energy and Minerals Graynolls Federal No. 1, sec. 22, T. 12 S., R. 21 E., Uintah County, Utah.
33. Phillips Petroleum Company Two Waters No. 1, sec. 22., T. 14 S., R. 25 E., Uintah County, Utah.
34. Continental Oil Company No. 22–1 Federal, sec. 22, T. 9 S., R. 20 E., Uintah County, Utah.
35. Phillips Petroleum Company Watson "B" No. 1, sec. 34, T. 9 S., R. 25 E., Uintah County, Utah.
36. Phillips Petroleum Company No. 1 Hells Canyon, sec. 12, T. 2 S., R. 104 W., Rio Blanco County, Colorado.
37. Superior Oil Company No. 1 Hellman–Government, sec. 5, T. 3 S., R. 101 W., Rio Blanco County, Colorado (Hallgarth, 1959).
38. Phillips Petroleum Company No. 15 Douglas Creek Arch, sec. 14, T. 1 S., R. 102 W., Rio Blanco County, Colorado.
39. The Texas Company No. 70–32 UPRR, sec. 32, T. 2 N., R. 102 W., Rio Blanco County Colorado (Hallgarth, 1959).
40. The Texas Company and the California Company No. 20 Government, sec. 34, T. 3 N., R. 94 W., Rio Blanco County, Colorado (Hallgarth, 1959).
41. Stanolind Oil and Gas Company Scott No. 1, sec. 20, T. 1 N., R. 93 W., Rio Blanco County, Colorado.
42. Buford Oil Company Government No. 1, sec. 16, T. 1 N., R. 91 W., Rio Blanco County, Colorado.
43. Phillips Petroleum Company No. 1 Linville Government, sec. 1, T. 1 N., R. 92 W., Rio Blanco County, Colorado.
44. Benedum Trees Oil Company Daugherty Government No. 1, sec. 7, T. 1 N., R. 88 W., Garfield County, Colorado.
45. Phillips Petroleum Company Poose Creek No. 1, sec. 10, T. 2 N., R. 88 W., Rio Blanco County, Colorado.
46. Sam Gary No. 11–15 Aspegren, sec. 2, T. 1 S., R. 85 W., Routt County, Colorado.
47. Miami Oil Producers No. 1 O'Brien, sec. 14, T. 4 N., R, 90 W., Moffat County, Colorado.
48. Texaco Oil Company No. 7 Government Treleaven, sec. 31, T. 5 N., R. 90 W., Moffat County, Colorado.
49. Stanolind Oil and Gas Company No. 24 Parkinson Government, sec. 22, T. 4 N., R. 92 W., Moffat County, Colorado (Hallgarth, 1959).
50. Stanolind Oil and Gas Company Blue No. 1, sec. 35, T. 6 N., R. 96 W., Moffat County, Colorado (Hallgarth, 1959).
51. Humble Oil Company No. 1 Lay Creek, sec. 13, T. 8 N., R. 93 W., Moffat County, Colorado.
52. Gulf Energy and Minerals Company No. 1 Slater Creek, sec. 7, T. 11 N., R. 89 W., Moffat County, Colorado.
53. Phillips Petroleum Company No. 8 Baggs Unit, sec. 10, T. 12 N., R. 92 W., Carbon County, Wyoming.

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Miscellaneous Field Studies Maps are multicolor or black-and-white maps on topographic or planimetric bases on quadrangle or irregular areas at various scales. Pre-1971 maps show bedrock geology in relation to specific mining or mineral-deposit problems; post-1971 maps are primarily black-and-white maps on various subjects such as environmental studies or wilderness mineral investigations.

Hydrologic Investigations Atlases are multicolored or black-and-white maps on topographic or planimetric bases presenting a wide range of geohydrologic data of both regular and irregular areas; the principal scale is 1:24,000, and regional studies are at 1:250,000 scale or smaller.

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