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Chapter 2

Structural evolution of piggyback basins in the Wyoming-Idaho-Utah thrust belt

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ABSTRACT

The depositional histories of three piggyback basins are related to the surface and subsurface thrust geometries that controlled early Eocene basin development during the final phase of shortening in the Idaho-Wyoming-Utah thrust belt. Early Eocene strata of Bear Lake Plateau basin, Fossil Basin, and the LaBarge basin occupy similar structural positions on the Absaroka, Crawford, and Hogsback thrust sheets respectively. The basin depocenters formed above flat décollement surfaces of underlying thrust sheets. Trailing basin margins formed above the trailing footwall ramps of the thrust sheets that immediately underlie the basins. Leading basin margins formed along the leading edges of thrust sheets that are underlain by the ramps of structurally lower thrusts. Sedimentary facies distributions and sediment dispersal patterns were controlled by uplift of the basin margins rather than by subsidence of depocenters. Periods of basin margin uplift are identified by the presence of alluvial fan sediments shed from the uplifted margins, patterns of onlap and overstep along and across the margins, and cross-cutting structural relations of folded or faulted early Eocene strata along the basin margins. Uplift was accomplished through reactivation of slip along the trailing footwall ramps of the Crawford and Absaroka thrusts, late slip on the Hogsback trailing ramp, and slip on the LaBarge and Calpet thrusts in the Hogsback footwall. The late thrust slip in the interior of the thrust belt was a type of break back, or out-of-sequence thrust slip, that deviated from the previous foreland-directed sequence of thrusting.

Reactivated thrust uplift and footwall uplift are the two uplift processes common to the three basins studied in this report. Reactivated thrust uplift was generally confined to the trailing ramp areas of thrust sheets, where slip along the trailing ramp produced ramp-rooted imbricate thrusts and fault-propagation folds. Footwall uplift was confined to the leading basin margins, where the basin margins and the leading edges of the thrust sheets beneath them were translated up the ramps of structurally lower thrusts. The reactivation of thrusts beneath piggyback basins across the width of the Wyoming-Idaho-Utah thrust belt is attributed to impeded slip along the frontal thrusts of the belt. Slip impedance along the thrust front is attributed to the interaction of four processes: buttressing of the frontal thrusts by coeval foreland basement uplifts, fault deflection above preexisting foreland basement warps, decrease in regional wedge taper from the combined effects of foreland subsidence and thrust belt erosion, and rheologic changes related to stratigraphic changes toward the foreland.

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INTRODUCTION

The southern half of the Wyoming-Idaho-Utah thrust belt is covered by early Eocene alluvial, fluvial, and lacustrine strata of the Wasatch and Green River formations that were deposited in a series of basins separated by discrete thrust-related structures during the final phase of regional thrust shortening. The basins were transported eastward throughout their development in a "piggyback" manner above the regional sole décollement. These piggyback basins (Ori and Friend, 1984) overlie individual thrust sheets that were emplaced and deeply eroded prior to basin development. The basin margins were locally sites of uplift above the trailing ramps and leading edges of thrust sheets. The early Eocene piggyback basins record a period of slip reactivation on long dormant faults in the interior of the belt that marked a change from the previous sequence of foreland-directed thrusting. The internal reactivated fault segments formed break-back thrusts behind the regional thrust front.

The purposes of this chapter are to (1) illustrate the thrust geometries and uplift processes that localized piggyback basin margins, (2) show the sedimentologic responses of individual basins to the growth of basin margin structures, and (3) provide mechanical explanations for the change in the style and sequence of thrusting that was responsible for the development of early Eocene piggyback basins.

The interpretation of break-back thrusting is a recent development in studies of the Wyoming-Idaho-Utah thrust belt. Regional studies (Armstrong and Oriel, 1965; Royse and others, 1975; Wiltschko and Dorr, 1983) present a clear record of the overall foreland-younging thrust sequence through the belt. The clearest example of a break-back thrust sequence was presented by Royse (1985) and Hunter (1988) for the Prospect thrust system in the northern Wyoming thrust belt where cross-cutting relationships and tectogenic sediments constrain thrust timing. Similar interpretations have been presented for the northern Darby (Dixon, 1982) and Absaroka (McBride and Dolberg, 1990) thrusts based on structural style. However, the lack of thrust-related sediments to date deformation permits alternative interpretations of the Darby (Jones, 1984) and Absaroka (Lageson, 1984; Woodward, 1986) thrust sequences. The piggyback basins of the central and southern thrust belt contain tectogenic sediments that are related to the subsurface structures defined by well and seismic data. The area is ideally suited for dating late thrust events, for defining the structural style of break-back thrusts, and for delineating the mechanical conditions responsible for the change in the regional thrust sequence during the last phase of thrust belt shortening.

Three basins are examined in this chapter: Bear Lake Plateau basin, which overlies the Crawford thrust sheet; Fossil Basin, which overlies the Absaroka thrust sheet; and the LaBarge basin, which overlies the Hogsback thrust sheet (Fig. 1). Both the Fossil Basin (Oriel and Tracey, 1970; Lamerson, 1982; Hurst and Steidtmann, 1986) and the LaBarge basin (Oriel, 1962, 1969; Blackstone, 1979; Dorr and Gingrich, 1980) have been studied in the past. This chapter provides the first detailed description of the structure, stratigraphy, and sedimentology of Bear Lake Plateau basin.

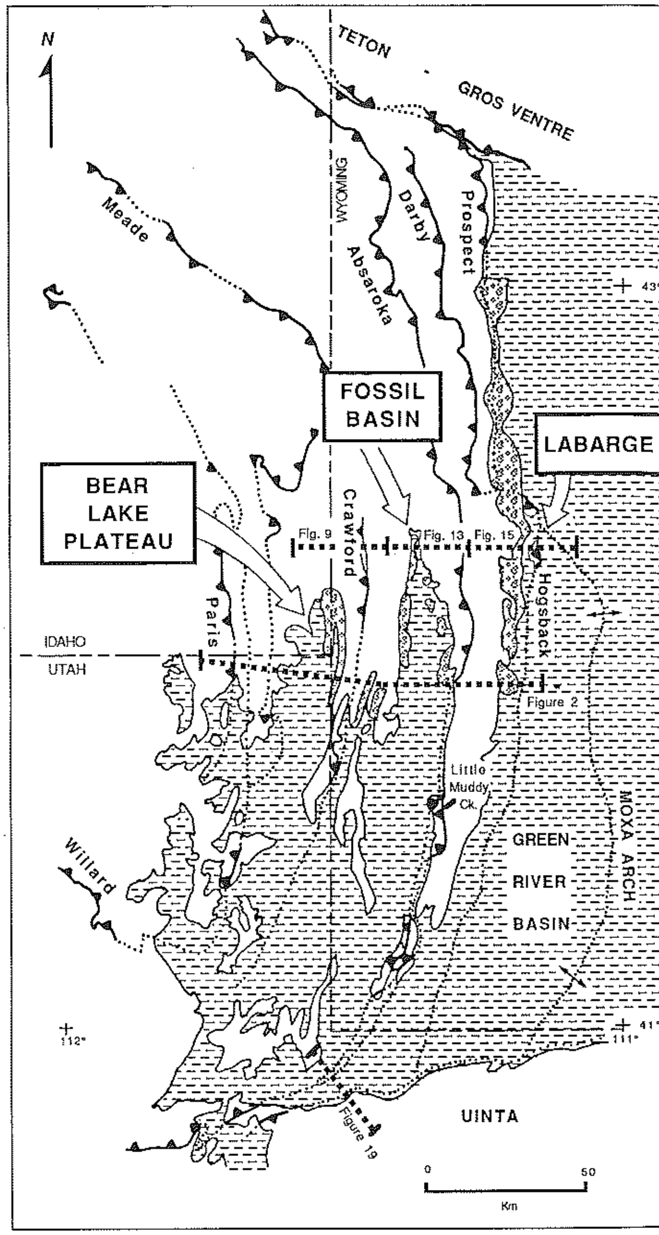


Figure 1. Index map of the Wyoming-Idaho-Utah thrust belt showing the location of principal thrust faults and foreland uplifts and their positions relative to the three early Eocene basins discussed here. The dash pattern represents fine-grained strata of the Wasatch and Green River formations. The stipple pattern represents diamictites of the Wasatch Formation (modified and generalized from Oriel and Platt, 1980; and Lamerson, 1982). Heavy dashed lines indicate cross sections shown in Figures 2, 9, 13, 15, and 19. Thrust faults, shown with solid lines where exposed and dotted lines where buried, have sawteeth on the hanging wall.

The method of investigation is to: (1) briefly describe the common structural elements of the piggyback basins and their respective thrust sheets; (2) provide geometric models to explain the structural similarities between basins; (3) relate the geometric models to the sedimentologic and structural features of Bear Lake Plateau basin, northern Fossil Basin, and the LaBarge basin; and finally (4) provide an explanation for the late deformation that caused segmentation of the top of the regional thrust wedge during the last phase of shortening across the Wyoming-Idaho-Utah thrust belt.

STRUCTURAL SETTING OF PIGGYBACK BASINS

Figure 2 illustrates several similarities between different piggyback basins. Angular unconformities between early Eocene basin sediments and Mesozoic and Paleozoic strata of underlying thrust sheets demonstrate that the basins formed well after the main-phase emplacement of their respective thrust sheets. Main-phase slip of the Crawford thrust was assigned a Coniacian-Santonian age based on the structural position and age of the Echo Canyon Conglomerate in the footwall of the southern Crawford thrust (Royse and others, 1975). This assignment is somewhat ambiguous because the precise southern trace of the Crawford thrust is concealed by Maastrichtian through early Eocene strata. Recently, DeCelles (1988) reported that the clasts in the Echo Canyon Conglomerate were probably derived from multiple sources, including the initial uplift of the basement-cored anticlinorium of the Wasatch Range carried above the Crawford thrust. By either interpretation, the Crawford thrust was emplaced by Coniacian-Santonian time. Main emplacement of the Absaroka thrust sheet is constrained as Santonian to Campanian in age by stratigraphic bracketing (Royse and others, 1975), with minor reactivation during the late Maastrichtian or early Paleocene (Lamerson, 1982). Final slip on the Hogsback thrust is stratigraphically bracketed to late Paleocene (Warner and Royse, 1987), with early Eocene footwall imbrication (Dorr and Gingrich, 1980). These timing constraints indicate that the early Eocene piggyback basins formed approximately 26 m.y. after emplacement of the Crawford thrust sheet and 16 m.y. after main-phase slip on the Absaroka thrust. The basins formed 6 m.y. after the final minor slip along the Absaroka thrust trace and immediately following slip along the Hobsback thrust trace. Piggyback basin sedimentation was concurrent with final slip on the Hogsback footwall imbricates, but it also postdates final slip on these faults.

The basins were also translated eastward during their development by slip on the underlying regional sole décollement. The basins were carried "piggyback" (Ori and Friend, 1984) above the underlying décollement during late slip on the Hogsback thrust and its footwall imbricates at the leading edge of the thrust belt.

The piggyback basins also exhibit similar patterns of deposition. The main basin depocenters lie above footwall décollements in the underlying thrust planes where thrust surfaces are parallel to bedding in low shear strength materials—such as the Jurassic salt décollement of the Crawford thrust and the Cretaceous shale décollements of the Absaroka and Hogsback thrusts. Trailing basin margins were the sites of uplift above trailing footwall ramps (position A on Figure 2), where the thrust surfaces cut upward through bedding in higher strength Paleozoic and Mesozoic units. Leading basin margins formed where the fronts of the thrust sheets were uplifted above trailing footwall ramps of structurally lower thrust sheets (position B on Figure 2).

Figure 2 also illustrates that basin deposition was not controlled by discrete subsidence of the depocenters as proposed by Oriel and Tracey (1970, p. 33) and Rubey and others (1975, p. 14) for Fossil Basin. The lack of differential subsidence between basin margins and centers is demonstrated by the homoclinal dip of the basement surface immediately beneath the basins and across the margins as constrained by seismic, magnetic, and deep well data that have become available since 1975. This study demonstrates that rather than being caused by local subsidence, these basins were formed by uplift of erosionally resistant strata along the two basin margins within the larger subsiding region of the eastern thrust belt and the Green River Basin.

GEOMETRIC MODELS FOR BASIN MARGIN UPLIFT

Basin margin uplift is controlled by two processes illustrated on Figure 3. The first process is reactivated slip on the major thrust that immediately underlies a piggyback basin. The second process is termed footwall uplift, where the leading basin margin and the leading edge of the thrust plate beneath it are translated up the trailing ramp of a structurally lower thrust. Footwall uplift previously has been referred to as *passive uplift* (Steidtmann and Schmitt, 1988) because it does not require slip across the thrust plane that bounds the thrust sheet beneath the tectogenic deposits. Schmitt and Steidtmann (1990) more recently used *interior ramp supported uplift* to describe the same phenomenon. Both terms described the case in which a mechanically locked thrust sheet in the interior of a thrust belt is uplifted above the trailing footwall ramp of a younger thrust. Because this chapter discusses uplift in the deeper, interior parts of the thrust belt as well as shallower level uplift along the leading edge of the belt, the general term *footwall uplift* is used to describe uplift of a mechanically locked thrust plane by slip on an underlying thrust.

Figure 3 presents a series of scaled geometric models for basins generated by both reactivated thrust slip and footwall uplift. The models are drawn with simple ramp-flat geometries that generalize the common structural features of the Crawford, Absaroka, and Hogsback thrust sheets. Figure 3a represents main-phase emplacement of a thrust sheet, which is shown in Figure 3b as eroded to a near-planar topographic surface prior to the development of any piggyback basin. Local topographic relief resulting from differential erosion could influence early subaerial deposition at the site of the later piggyback basin, particularly where steeply dipping, resistant strata are exposed above the trailing ramp and leading edge of the thrust sheet.

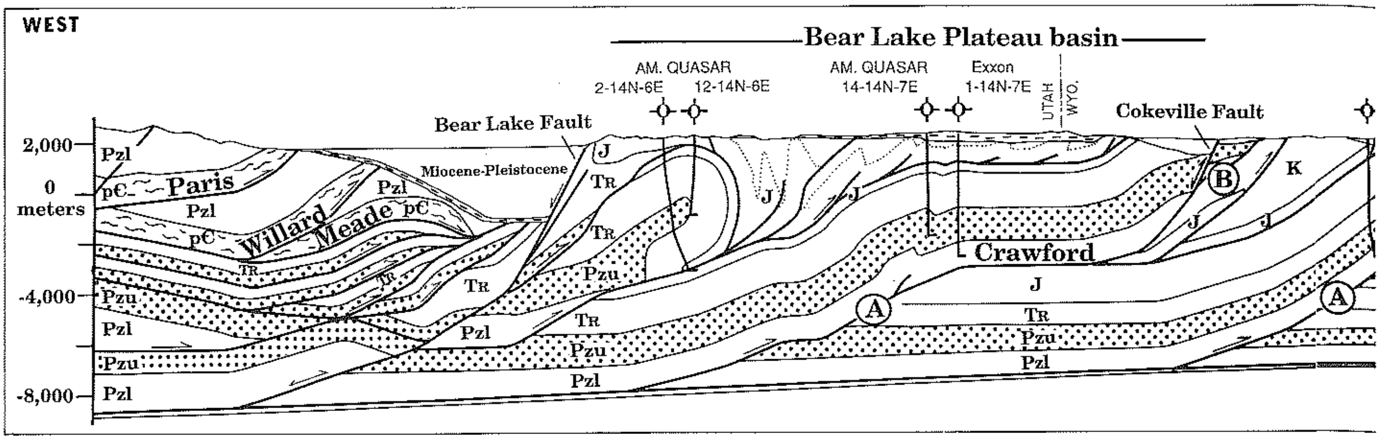


Figure 2. West-east cross section between the Paris and Hogsback thrust fronts showing the structural position of Bear Lake Plateau basin and Fossil Basin. See Figure 1 for location. Early Eocene basin fill is defined by dash pattern. The depocenters of Bear Lake Plateau basin and Fossil Basin overlie footwall décollements in the underlying Crawford and Absaroka thrust faults respectively. The trailing basin margins lie above the trailing footwall ramps (position A) of the underlying thrust sheets, and the leading basin margins lie above the leading edges of the underlying thrust sheets (position B). LaBarge basin is located 35 km (21 m) north of this line of section in the position of the western margin of the Green River Basin on the Hogsback thrust sheet. p€: Proterozoic, Pzl: early Paleozoic, Pzu: late Paleozoic, ₮: Triassic, J: Jurassic, K: Cretaceous rocks. Vertical scale = horizontal scale. The Absaroka and western Hogsback thrust sheets are modified from Lamerson (1982, Plate 6).

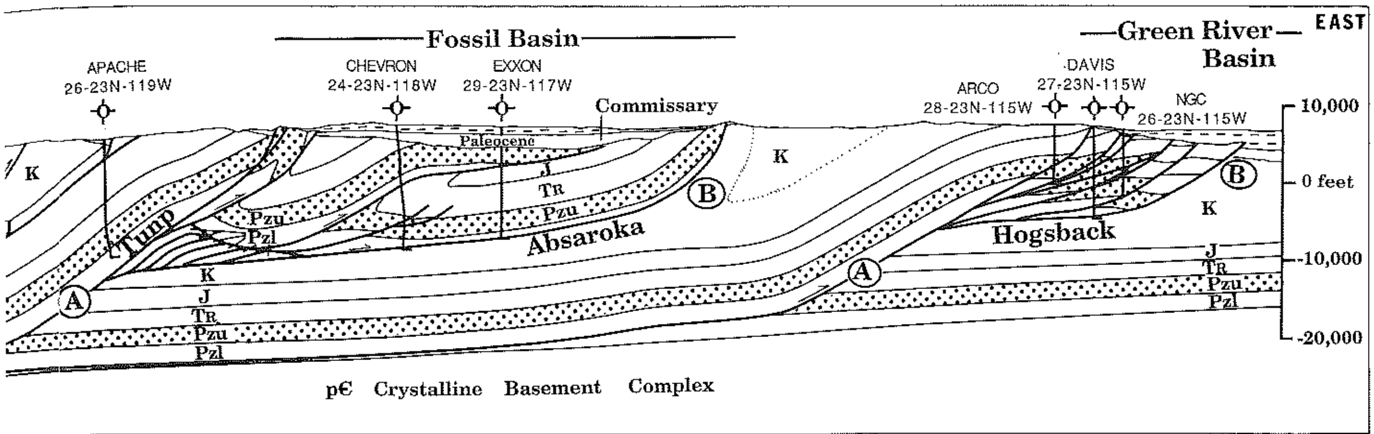
Figure 3c illustrates that reactivation of thrust slip on a previously locked thrust plane can affect either the entire thrust surface or just the trailing area of the thrust. In the first case, slip is transferred continuously from the sole décollement, up the trailing ramp, across the succeeding footwall décollement, and finally to the thrust front, where uplift is accommodated by either translation and rotation up a leading ramp or by imbrication at the thrust tip. Both the leading and trailing uplifts are source areas for clastic sediments, and the depression between them is the depocenter of the piggyback basin. Figure 3c also illustrates a second case, in which slip is confined to the trailing footwall ramp area. This slip produces a break-back thrust that bypasses the mechanically locked décollement of the thrust sheet.

Break-back thrusts and fault-propagation folds are typical late structures above trailing ramps. If such structures developed early in a foreland-directed sequence, then there should be locations along strike where the imbricate lies behind the ramp (Jones, 1984) or where it has been translated up onto the frontal décollement. Instead, these structures are consistently found immediately above the trailing ramp as a common structural motif in the Wyoming-Idaho-Utah thrust belt (Boyer, 1986, Fig. 5). Although kinematic arguments based on structural style can be ambiguous, tectogenic sediments permit independent dating of ramp-rooted structures along the trailing margins of piggyback basins.

Figure 3d illustrates a typical foreland-younging sequence of thrusting in which an early thrust sheet becomes mechanically locked, and further shortening is accommodated by propagation of a younger thrust beneath and in front of the locked thrust. The locked thrust becomes part of the hanging wall of the younger thrust and is translated toward the front of the belt by slip on the younger thrust plane. Figure 3e shows the older, locked thrust sheet translated with its footwall onto the trailing ramp of the younger thrust fault. The subsequent uplift and rotation of the older, locked thrust sheet is called footwall uplift with reference to the older, structurally higher, and mechanically locked thrust plane. Footwall uplift requires no slip on the older thrust plane. Footwall uplift can produce a leading basin margin in the hanging wall of the locked thrust, but footwall uplift does not produce the trailing basin margins seen in the Wyoming-Idaho-Utah thrust belt—where piggyback basins are bordered by uplifts on both basin margins.

Figure 3f illustrates that both reactivated thrust uplift and footwall uplift can act in unison to respectively create the trailing and leading margins of a piggyback basin. Both processes were operative in Bear Lake Plateau basin, Fossil Basin, and the LaBarge basin, although the relative timing of leading and trailing basin margin uplift differs for each basin.

Different reasons exist for the causes of uplift of trailing and leading basin margins for the simple ramp-décollement geometries illustrated in Figure 3. Uplift of the frontal basin margin can be the result of either thrust reactivation or footwall uplift, whereas uplift at the trailing basin margin provides unambiguous evidence for reactivated thrust slip.



METHODS FOR DATING BASIN MARGIN UPLIFT

Periods of basin margin uplift are identified by combined observations of (1) alluvial fan sediments shed from the basin margins, (2) patterns of onlap and overstep along and across the margins, and (3) crosscutting structural relations of folded or faulted piggyback basin strata along the basin margins.

The distribution of early Eocene alluvial fan deposits in the Wyoming-Idaho-Utah thrust belt is shown in Figure 1. These deposits have been generally mapped and discussed as diamictite or conglomerate facies of the Wasatch Formation, and they include the diamictite member of the Wasatch Formation in eastern Bear Lake Plateau (this chapter); Wasatch diamictite of Oriel and Platt, 1980), the Tunp Member of the Wasatch Formation in northern Fossil Basin (Oriel and Tracey, 1970; Rubey and others, 1975; Hurst and Steidtmann, 1986), and the conglomerate member (Oriel, 1969) or Lookout Mountain Conglomerate Member (Dorr and Gingrich, 1980) of the Wasatch Formation in the La Barge basin as shown on Figure 4. The diamictites are mainly matrix-supported gravels that contain clasts derived from proximal source areas of Mesozoic and Paleozoic rocks along the structurally high margins of the piggyback basins.

Both tectonics and climate have been used to explain the origin of Wasatch diamictites. This chapter highlights the tectonic setting of the diamictites, although a favorable climate certainly contributed to diamictite deposition. Oriel and Tracey (1970, p. 29) and Tracey and others (1961) considered diamictite deposition in the Wasatch Formation to have been primarily caused by steep paleotopography generated by differential erosion combined with the humid subtropical to savanna early Eocene climate of the middle Rocky Mountains (Leopold and MacGintie, 1972). The diamictites could have been generated by periodic but intense rainfall that induced gravitational flow in areas of steep paleotopography. Oriel and Tracey (1970), Rubey and others (1975), and Tracey and others (1961) did not recognize the importance of early Eocene thrust-related uplift within the Crawford and Absaroka thrust sheets and in the western part of the Hogsback thrust sheet. Thus, the paleotopographic relief discussed by these workers would have been inherited from the topography established at the end of main emplacement of these thrust sheets. Lamerson (1982) did recognize sites of early Eocene thrust-related uplift along the margins of Fossil Basin area that Hurst and Steidtmann (1986) correlated to source areas for Wasatch diamictites. Diamictites occur only in specific structural settings in each of the three basins discussed below.

Paleotopographic relief caused by differential erosion of thrust sheets was a contributing but not sufficient condition for diamictite development because examples of erosionally resistant strata beneath fine-grained deposits of basin depocenters are found in all three basins. Instead, the diamictite source areas along the basin margins were areas of rejuvenated paleotopography that were supported by early Eocene uplift.

Patterns of onlap and overstep of early Eocene strata onto and across the basin margins can be the result of infilling of remnant paleotopography or of progressive infilling of paleotopography supported by coeval uplift. Periods of uplift are identified in this chapter by sequential angular unconformities within the Wasatch Formation and between the Wasatch Formation and older Tertiary units.

Crosscutting relations between thrust faults, thrust-related folds, and early Eocene sediments provide a third criterion for dating periods of basin margin uplift. Recent subsurface data, combined with new mapping of the Bear Lake Plateau area and field reconnaissance of the Fossil Basin and LaBarge basin, provide greater resolution of the early Eocene deformation associated with structures mapped by Oriel (1969), Rubey and others (1975, 1980), and Oriel and Platt (1980) in the three basins.

BEAR LAKE PLATEAU BASIN

Structural setting

Bear Lake Plateau basin lies on the hanging wall of the Crawford thrust sheet near the point common to Wyoming, Idaho, and Utah (Fig. 1). Two structural elements of the thrust sheet bound early Eocene depositional areas that were affected by

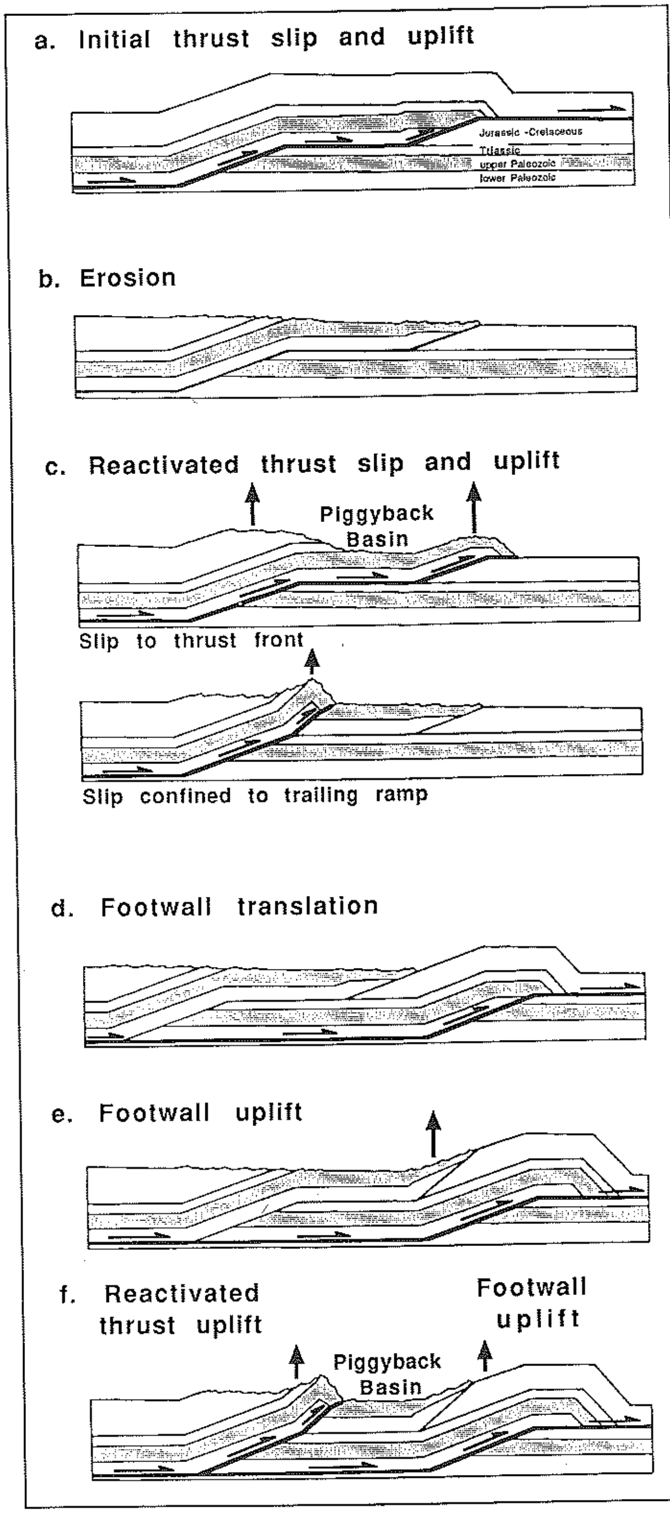


Figure 3. Scaled geometric models for basin margin uplifts that approximate the structural features common to the Crawford, Absaroka, and Hogsback thrust sheets. Thrust fault lengths undergoing slip in each diagram are shown by bold lines and arrows, whereas the other thrust faults are considered to be mechanically locked.

different sedimentation histories: (1) the leading edge of the Crawford thrust including Cokeville anticline (Fig. 5); and (2) the trailing ramp area of the Crawford thrust, which lies beneath Pegram anticline to the south and Sublette anticline to the north (Fig. 5). The thrust-related structure of Bear Lake Plateau has been modified by Miocene to Holocene normal faulting along the plateau margins. The Cokeville normal fault bounds the east side of a shallow half graben east of Bear Lake Plateau, and the Bear Lake normal fault separates the western border of the plateau from Bear Lake graben to the west (Fig. 5). Normal faults mapped by Rubey and others (1980) and Oriel and Platt (1980) on eastern Bear Lake Plateau were not found during remapping of area (Fig. 5).

Bear Lake Plateau basin is bound to the east by the leading edge of the Crawford thrust sheet. The Crawford thrust crops out east of Cokeville, Wyoming, but it is covered beneath Tertiary and Quaternary valley fill where it has been displaced by the Cokeville normal fault to the south (Fig. 5). The Stoffer Ridge thrust crops out east of Cokeville as a footwall imbricate to the Crawford thrust that branches from the frontal Crawford décollement in Jurassic salt. North of Cokeville, the leading edge of the Crawford thrust is located within a fold belt above the salt décollement (Coogan and Yonkee, 1985). Cokeville anticline is a hanging-wall ramp anticline that overlies this décollement (Coogan and Yonkee, (1985). The Wasatch Formation onlaps Mesozoic rocks of the leading edge of the Crawford sheet along the east side of Bear Lake Plateau basin where Wasatch diamictites have been rotated to westward dip above an angular unconformity.

The central part of Bear Lake Plateau basin is underlain by Pegram anticline, a fault-propagation fold that lies above the Crawford trailing ramp. The anticline is exposed along the northern margin of the plateau, and it has been drilled in the subsurface beneath the Wasatch Formation where its approximate location is dotted on Figure 5. Pegram anticline divides Bear Lake Plateau basin into two subbasins, with an early fluvial basin east of Pegram anticline and the depocenter of a late lacustrine basin to the west. Pegram anticline plunges north along the northern margin of Bear Lake Plateau, where the subsurface position of the Crawford trailing ramp shifts eastward across a lateral thrust ramp to where it underlies Sublette anticline in the northern part of Figure 5.

Mapping by Valenti (1982) and Dover (1985) along the southern edge of Bear Lake Valley demonstrates that the western boundary of Wasatch deposition extended west of Bear Lake Plateau prior to Miocene to Holocene normal faulting. The Wasatch Formation is continuous south of Bear Lake Plateau to where it onlapped Paleozoic strata of the Paris thrust hanging wall to the west (Fig. 1).

Age and stratigraphic position

Figure 4 summarizes the age and stratigraphic relatinships recognized for the Wasatch Formation and associated younger and older Tertiary deposits of Bear Lake Plateau, and Sublette

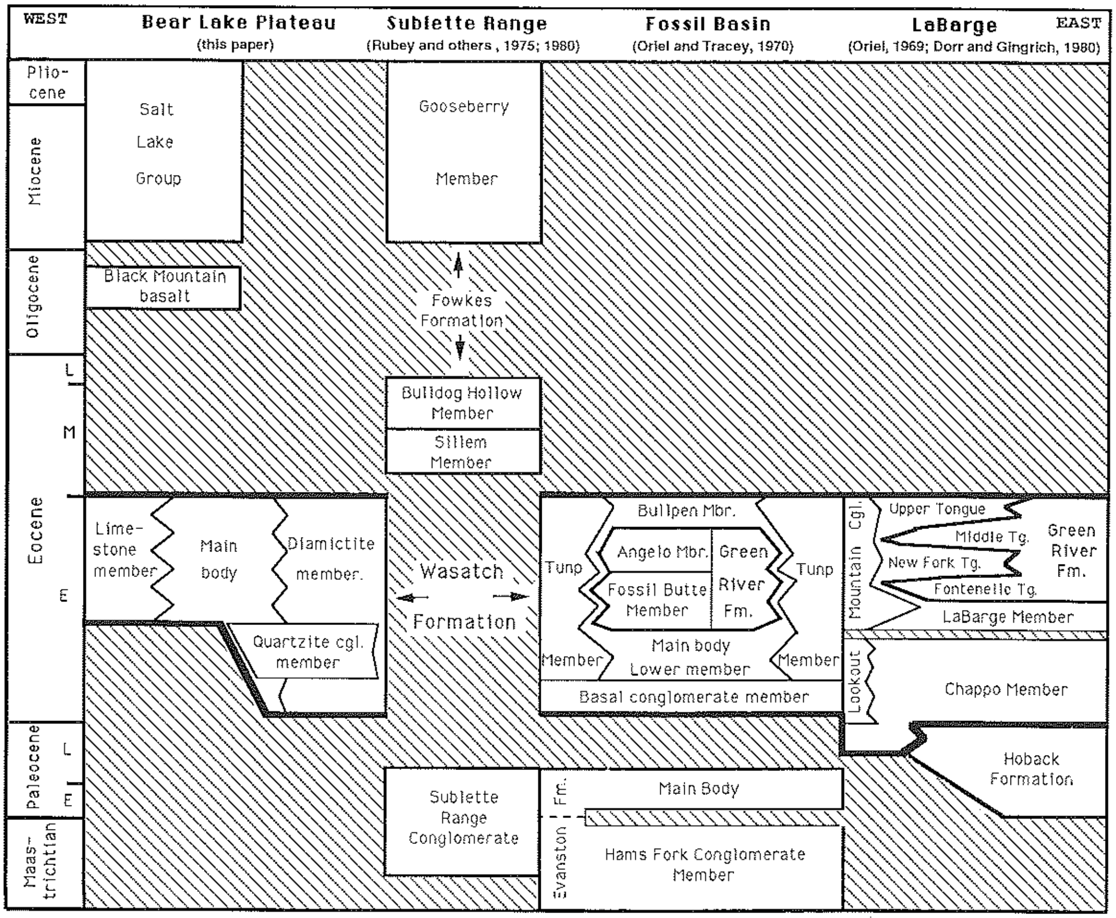


Figure 4. Latest Cretaceous and Tertiary stratigraphic relationships for the Bear Lake Plateau basin, Sublette Range, northern Fossil Basin, and LaBarge basins.

Range, Fossil Basin, LaBarge, and surrounding areas. Mansfield (1927) initially mapped most of Bear Lake Plateau as Eocene Wasatch Formation based on surface correlation to dated Wasatch deposits in other areas. Oriel and Platt (1980) subsequently mapped most of Bear Lake Plateau north of the Utah border as Pliocene and Miocene Salt Lake Formation based on a collection of *Leporidae* teeth from a white, tuffaceous conglomorate in the northwestern corner of Bear Lake Plateau (Oriel and Tracey, 1970, p. 37). Oriel and Platt (1980) also mapped thin outcrop belts of Wasatch diamictite and Maastrichtian-Paleocene Evanston Formation along the eastern margin of Bear Lake Plateau. The diamictite of Oriel and Platt (1980) is informally designated the diamictite member of the Wasatch Formation in this chapter (Figs. 4 and 5), whereas conglomerates mapped as the Evanston Formation by Oriel and Platt (1980) are here incorporated into the quartzite conglomerate member of the Wasatch Formation based on intertonguing relationships with the diamictite member.

Remapping of the area (Fig. 5) revealed three overlapping relationships for the Bear Lake Plateau basin. (1) The youngest overlapping beds are the Salt Lake strata sampled by Oriel and Tracey (1970), which are restricted to small debris flow deposits that unconformably overlie the Wasatch strata in northeastern Bear Lake Plateau (Fig. 5). (2) The Wasatch Formation is also overlain and crosscut by an olivine basalt on Black Mountain, Utah, shown in the western part of Figure 5. The basalt yields an Oligocene K-Ar whole rock age of 28.8 ± 1.7 m.y. (Mark Jensen, Utah Geological and Mineral Survey, written communication, 1989). (3) Finally, the Wasatch Formation is overlain in angular unconformity by tuffaceous beds of the middle Eocene Sillem Member of the Fowkes Formation along the eastern margin of Bear Lake Plateau (Fig. 5). Rubey and others (1980) interpreted the Fowkes-Wasatch contact as a normal fault, but the new mapping reveals that the Fowkes Formation dips 15° east above and adjacent to 15 to 25° west-dipping Wasatch Formation. This overlapping relationship provides a pre–middle Eocene upper age limit for the Wasatch Formation of Bear Lake Plateau. Within the Wasatch Formation, early Eocene gastropods occur in limestones that intertongue with the upper part of the Wasatch clastic sequence in western Bear lake Plateau basin (H. P. Buchheim, Loma Linda University, personal communication, 1989).

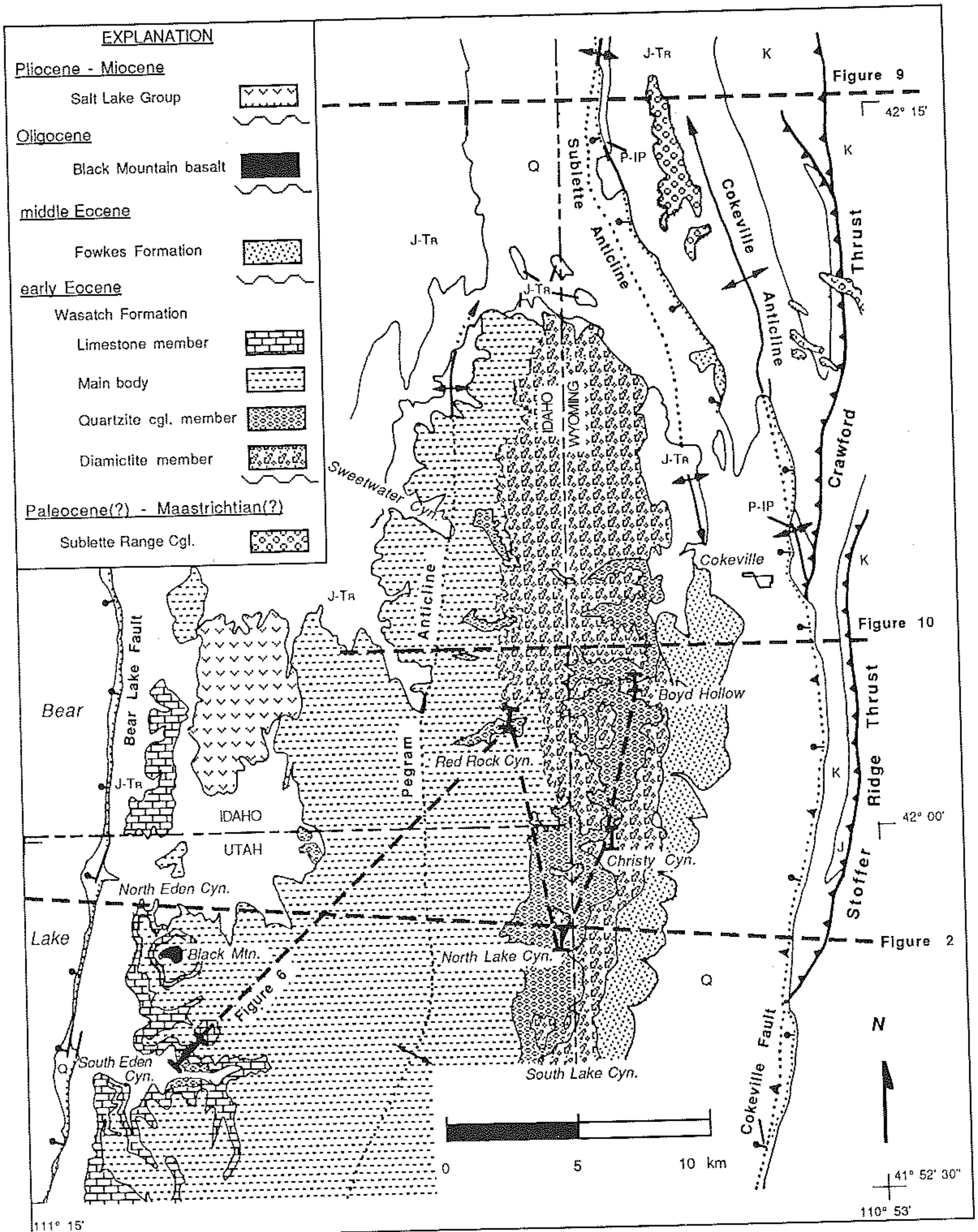


Figure 5. Geologic map of Tertiary strata in the Bear Lake Plateau area, Wyoming-Idaho-Utah, showing principal surface folds and faults of the Crawford thrust sheet. Thrust faults have sawteeth on the hanging wall; normal faults have ball on hanging wall. Faults and folds are dotted where they are buried by Tertiary and younger strata. Members of the Wasatch Formation become progressively younger westward. The Salt Lake Group, Black Mountain basalt, and Fowkes Formation unconformably overlie the Wasatch Formation. The locations of structural cross sections are shown for Figures 9, 10, and 2. The locations of measured sections are shown along the line of Figure 6. P-IP = Permian and Pennsylvanian rocks; J-TR = Jurassic and Triassic rocks; J = Jurassic rocks; K = Cretaceous rocks; Q = Quaternary rocks.

The Sublette Range Conglomerate crops out along the west flanks of Sublette and Cokeville anticlines and above the Crawford thrust trace (Fig. 5). The conglomerate is undated, but it is placed beneath the Wasatch Formation on Figure 4 based on lithologic, sedimentologic, and provenance correlations to the Paleocene-Maastrichtian Evanston Formation in adjacent areas (Salat, 1989). Like the Evanston Formation, the Sublette Range Conglomerate has abundant red quartzite boulders and cobbles that have a unique provenance in the Lake Proterozoic section of the Willard and Paris thrust sheets. Furthermore, clasts derived from the Sublette Range Conglomerate are interpreted below to have been recycled into coarse-grained Wasatch strata along the eastern margin of Bear Lake Plateau basin.

Stratigraphy and sedimentology of the Wasatch Formation

Five measured sections constrain the early Eocene stratigraphy of Bear Lake Plateau basin: Christy Canyon and Boyd Hollow on the east side of the basin, North Lake Canyon and Red Rock Canyon in the central basin, and South Eden Canyon on the west flank of Bear Lake Plateau basin (Figs. 5 and 6). Four informal members of the Wasatch Formation were measured and mapped from east to west. In ascending stratigraphic order, the members include: (1) the diamictite member, (2) the quartzite conglomerate member, (3) the main body, and (4) the limestone member. Detailed measured sections of the five members are provided in Coogan (1992). A generalized cross section through the measured sections is shown on Figure 6, in which the relative stratigraphic position of each measured section is constrained by surface and photogeologic mapping of key beds between the sections.

Diamictite member. The diamictite member is best exposed in Christy Canyon and Boyd Hollow along the eastern margin of Bear Lake Plateau basin (Figs. 5 and 6). The diamictite member consists of debris- and mud-flow deposits that were transported westward and southwestward as part of an alluvial fan system shed from the Crawford thrust front and Sublette anticline. This interpretation is based on analysis of dominant lithofacies, thickness trends, lateral relationships, and clast provenance.

The diamictite member contains three major lithofacies: matrix-supported gravel, massive clast-supported gravel, and massive mudstone. The matrix-supported gravels are moderately organized where they have sand and silt matrix and display inverse grading and faint horizontal stratification that are indicative of deposition in noncohesive debris flows (Nemec and Steel, 1984). In contrast, clay-rich matrix-supported gravels (Fig. 7) display no sorting trends or stratification and probably formed as cohesive debris flows (Nemec and Steel, 1984). Disorganized, massive, clast-supported gravels are interpreted as having formed as rapidly deposited, clay-poor, noncohesive debris flows (Nemec and Steel, 1984; DeCelles and others, 1987). Massive mudstones

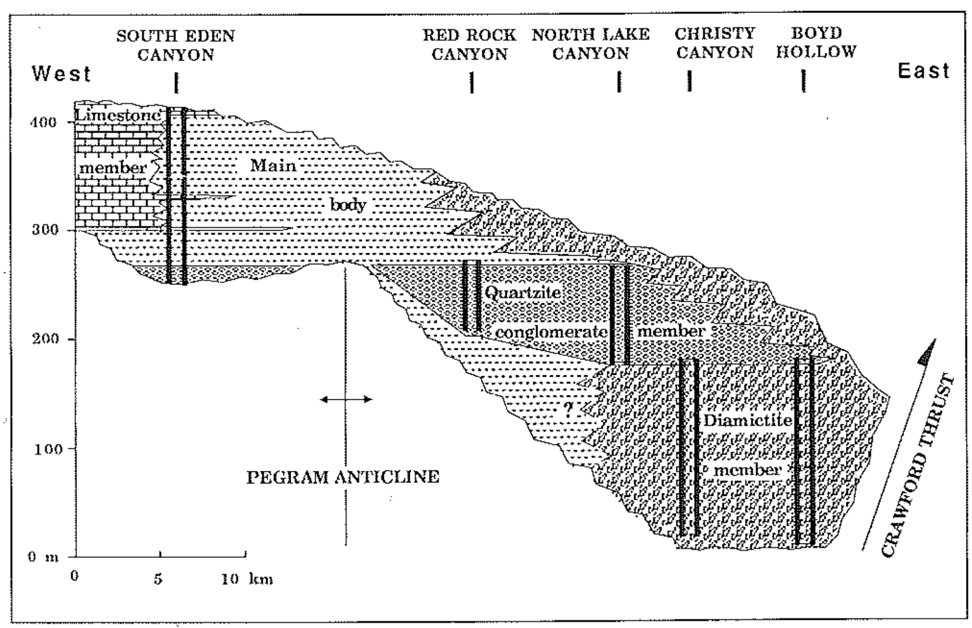


Figure 6. Stratigraphic correlation between informal members of the Wasatch Formation showing the position of underlying structures in the Crawford thrust sheet, the intertonguing relationships mapped on Figure 5, and the five measured sections (Coogan, 1992). The basin depocenter migrated through time from an early, fluvial depocenter in the quartzite conglomerate member of eastern Bear Lake Plateau basin to a late, lacustrine depocenter in western Bear Lake Plateau basin. The migration of the depocenter resulted from stratigraphic onlap and overstep across the crest of Pegram anticline.

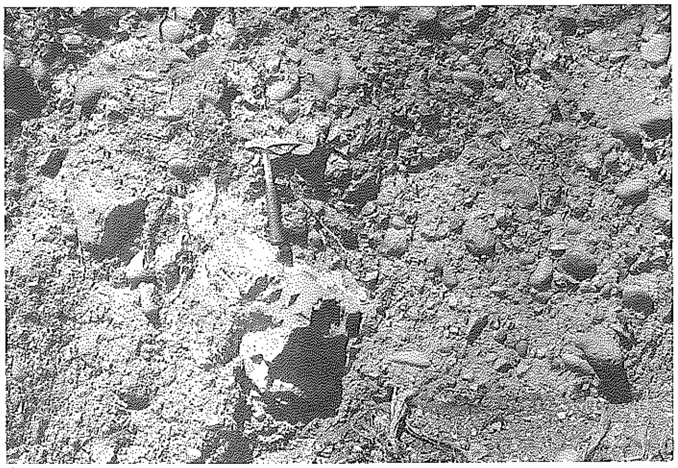


Figure 7. Typical outcrop of disorganized matrix-supported gravel of the diamictite member of the Wasatch Formation. Notice the poor sorting of both grain size and shape. The base of the hammer rests on an angular block of Nugget Sandstone that was probably transported a short distance from the source area in the Sublette Range or along the Crawford thrust front. The point of the hammer touches a well-rounded clast of Proterozoic quartzite that probably experienced an early episode of fluvial transport prior to its deposition in the Sublette Range conglomerate as well as a second period of debris-flow transport during which it was recycled into the diamictite member during uplift of the Sublette Range and the Crawford thrust front.

locally contain widely spaced matrix-supported boulders, cobbles, and pebbles that were probably entrained in cohesive mudflows. The noncohesive debris flows and cohesive debris- and mud-flow deposits of the diamictite member are characteristic of a proximal alluvial fan depositional environment. Minor horizontally stratified and trough cross-stratified sandstone that is interbedded with the gravel and mudstone probably formed as minor sheetflood and channel deposits that are commonly found on the surface of alluvial fans (Rust and Koster, 1984).

Thickness trends and the map distribution of the diamictite member demonstrate that the alluvial fan complex was shed from a source area immediately to the east and northeast in the vicinity of the Sublette Range and the Crawford thrust front. The diamictite member thins westward and southwestward and intertongues with both the braided stream deposits of the quartzite conglomerate member and the fluvial channel and floodplain deposits of the main body of the Wasatch to the west and southwest (Figs. 5 and 6). The provenance of clasts from the diamictite member also supports the interpretation of a proximal source area to the east and northeast.

Clast counts through the vertical sections of the diamictite member at Boyd Hollow and Christy Canyon exhibit a subtle inverse stratigraphy that was likely caused by unroofing of the sublette Range and the Crawford thrust front during the early Eocene (Coogan, 1992). The stratigraphy of these potential source areas is exposed along the east flanks of Sublette and Cokeville anticlines and astride the Crawford thrust to the east where the leading edge of the Crawford thrust sheet has not been displaced by the Cokeville normal fault. The Crawford thrust and Sublette and Cokeville anticlines are overlain with angular unconformity by erosional remnants of the Sublette Range Conglomerate (Fig. 5), which contains a diagnostic assemblage of well-rounded Late Proterozoic and Paleozoic quartzite clasts as well as abundant Paleozoic chert clasts (Salat, 1989). Highly folded lower Cretaceous through Jurassic rocks lie beneath the unconformity, and Triassic and Permo-Pennsylvanian rocks have been exhumed at deeper structural levels in Sublette and Cokeville anticlines. Within the diamictite member, the well-rounded, Late Proterozoic quartzite clasts and Paleozoic quartzite and chert clasts are concentrated at the base of the unit, whereas clasts of Jurassic carbonates and the Jurassic Nugget Sandstone are rare, and Triassic clasts are absent. In contrast, the upper part of the diamictite member contains abundant angular Jurassic clasts as well as the first appearance of Triassic clasts, but well-rounded chert and quartzite clasts are less common, and Proterozoic quartzite clasts are absent in the uppermost beds. The most plausible interpretation for these clast distributions is that the stratigraphically higher diamictite beds record progressive erosion into Sublette anticline and the frontal Crawford thrust structures from the level of the Sublette Range Conglomerate that caps the anticline to the lower Triassic level in the core of the anticline.

Quartzite conglomerate member. The quartzite conglomerate member occupies a north-south trending trough in eastern Bear Lake Plateau basin, with the most complete exposures located in North Lake and Red Rock canyons (Figs. 5 and 6). The quartzite conglomerate member was deposited in a south-flowing, gravel-dominated braided stream system that reworked the toes of the diamictite member alluvial fans. This interpretation is supported by lithofacies analysis, paleocurrent data, and lateral relationships with adjacent members.

The main gravel lithofacies of the quartzite conglomerate member is massive, clast-supported gravel with up to 80% well-rounded quartzite and chert clasts. The sheetlike massive gravels grade upward into trough cross-stratified gravels and sandstones. The massive gravels contain crude horizontal stratification and imbricate cobble trains. The gravel sheets were probably deposited as individual gravel bars in a proximal, gravel-dominated, braided stream system, based on their similarity to recent and ancient examples (Boothroyd and Ashley, 1975; Rust and Koster, 1984; Miall, 1978). The trough cross-stratified gravels and sandstones probably filled mixed sand and gravel channel systems that modified the bar tops after periods of gravel-dominated deposition (Miall, 1978).

Paleocurrent data support a low-sinuosity braided stream interpretation for the quartzite conglomerate member. The individual and summed paleocurrent plots for the c-axes of imbricated pebbles in Figure 8 show an axial drainage direction subparallel to the Wasatch facies transitions and the thrust-related structural trends of Bear Lake Plateau. The paleocurrents are directionally bimodal, with a principal drainage direction to the southwest and a second minor component to the southeast.

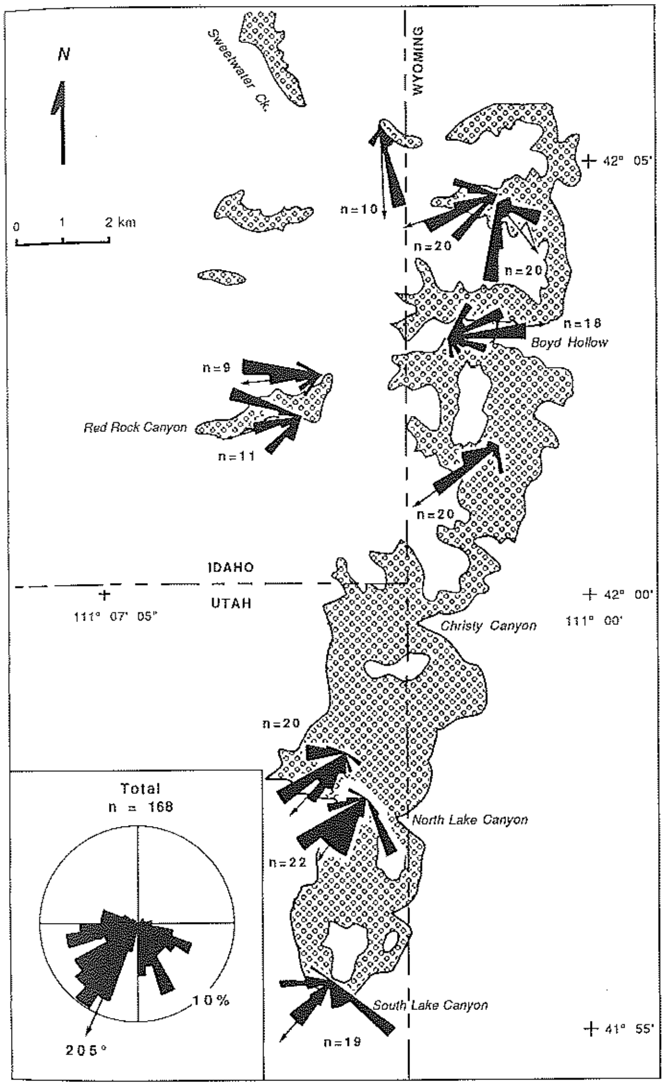


Figure 8. Rose diagrams of c-axis orientations for nonspherical cobbles in the quartzite conglomerate member of the Wasatch Formation. The vector mean for all measurements is 205°, showing a general southwestward current direction subparallel to the trend of Crawford hanging-wall structures and longitudinal to the basin axis.

The map distribution of the quartzite conglomerate member (Figs. 5 and 6) indicates that the braided alluvial plain was mainly limited to eastern Bear Lake Plateau between the diamictite member depositional front to the east and Pegram anticline to the west. Immediately west of the Red Rock Canyon section, the quartzite conglomerate member thins markedly toward the east flank of Pegram anticline, a phenomenon also observed at Sweetwater Canyon to the north (Fig. 5). Two isolated outcrops of the quartzite conglomerate member occur along the west flank of Pegram anticline overlying the basal angular unconformity above Jurassic rocks in North and South Eden canyons (Figs. 5 and 6). The conglomerate thins eastward in both locations, where it is truncated and overlapped by the main body of the Wasatch Formation. Pegram anticline was a paleotopographic high that separated the quartzite conglomerate member into eastern and western depositional areas.

Main body. The main body of the Wasatch Formation is a sand- and mud-dominated fluvial sequence that is best exposed in the South Eden section where it interfingers with the limestone member (Figs. 5 and 6). Red mudstone is volumetrically the most significant lithofacies of the main body. These mudstones are structureless wherever exposed, but they are interpreted to be floodplain muds and silts based on their relationship to laterally correlative trough cross-stratified sandstones. The sandstones are thin and laterally discontinuous with abrupt lower contacts on underlying mudstones. The bases of trough cross-stratified units often contain pebble lags that grade upward into sandstone. The lenticular shape, cross-bedding, and limited lateral extent of the trough cross-stratified sandstones lead to an interpretation of these units as channel sequences in low-sinuosity stream systems (Collinson, 1978). Horizontally laminated sandstones occur in the main body as very thin, tabular beds that occur within mudstones as well as within and above trough cross-stratified sandstones. The horizontally laminated sandstones may represent broad, shallow channel fills or possibly local crevasse splay deposits adjacent to larger channels (Collinson, 1978).

Sandstones in the main body of the Wasatch member were probably deposited in a sand- and mud-dominated fluvial system of individual short-lived stream channels bordered by mudstone deposition in interchannel areas. The main body fluvial deposits represent an intermediate transport system between the clastic source area of the diamictite member to the east and the lacustrine depocenter of the limestone member of the Wasatch Formation (Figs. 5 and 6).

Limestone member. The limestone member of the Wasatch Formation occupies a shallow lacustrine depocenter in western Bear Lake Plateau basin that is best exposed in the South Eden Canyon area (Figs. 5 and 6). The limestones contain abundant coarse clasts including oncolites, coated grains, and algal rip-up clasts. The large volume of pebbles within the limestone as oncolite nucleii and as coated grains implies that coarse material was continually supplied from the surrounding fluvial system.

The limestone member interfingers with the main body throughout the western part of Bear Lake Plateau, and it thickens dramatically to the southwest, where it is over 60 m (200 ft) thick in a continuous vertical sequence near the southeast corner of Bear lake (Figs. 5 and 6). The east to west transitions from the diamictite member alluvial fan deposits, through the main body fluvial system, and finally into the limestone member lacustrine depocenter imply a general westward or southwestward sediment transport direction during the later stages of deposition in the Bear Lake Plateau basin (Figs. 5 and 6). The locus of deposition and modes of transport contrast with the older depositional system of the quartzite conglomerate member in eastern Bear Lake Plateau. The change from the eastern to western depocenters through time was controlled by an interplay between infilling of structurally defined paleotopography and shifting sites of tectonic uplift.

Structural evolution of Bear Lake Plateau basin

Sublette anticline, Pegram anticline, and the Crawford thrust front were structural and paleotopographic highs along the margins of Bear Lake Plateau basin that served as source areas for clastic sediments. Surface and subsurface structural information is combined with stratigraphic data to establish the age and sites of uplift in the three areas.

Uplift of Sublette anticline. The clast provenance, paleocurrent directions, and stratigraphic relationships of the Wasatch Formation all indicate that Sublette anticline was a topographic high during early Eocene sedimentation in Bear Lake Plateau basin. The inverse stratigraphy of clasts in the diamictite member indicates that the Sublette Range was erosionally exhumed from Jurassic through lower Triassic levels during the early Eocene. Such downcutting requires uplift of the Sublette Range because all well and geophysical data indicate that there was no local subsidence of basement beneath the adjacent piggyback basins. Uplift is also necessary to explain the present elevation of the Sublette Range Conglomerate above and east of Sublette anticline. The Sublette Range Conglomerate crops out with approximately 10° southeast dip at elevations between 2,500 and 2,760 m (8,200 to 9,050 ft) along the east flank of Sublette anticline. Assuming that the Sublette Range Conglomerate is older than the Wasatch Formation, as indicated by Salat (1989) and this chapter, a minimum 625 m (2,050 ft) of uplift is needed to explain the elevation difference between the conglomerate and the base of the Wasatch Formation, which lies at elevations of 1,794 to 1,875 m (5,885 to 6,150 ft) in wells in the interior of eastern Bear Lake Plateau basin. The elevation difference cannot be attributed to Miocene to Holocene normal slip of the Cokeville listric normal fault because vertical throw on the fault is minimal where it soles into the footwall décollement of the Crawford thrust plane beneath the interior of the eastern basin. Similarly, a minimum of 305 m (1,000 ft) of uplift is required to explain the elevation difference between the Sublette Range Conglomerate above Sublette anticline and eastern outcrops of the conglomerate that lie at elevations between 2,195 m (7,200 ft) and 2,010 m (6,600 ft) along the east flank of Cokeville anticline and above the Crawford thrust trace (Fig. 5). A kinematic model for differential uplift of Sublette anticline relative to Bear Lake Plateau basin, Cokeville anticline, and the Crawford thrust front is developed below.

The present geometry of Sublette anticline and the structural position of the Sublette Range Conglomerate are shown in Figure 9a. Sublette anticline overlies the trailing ramp of the Crawford thrust where the ramp is constrained by the Pan Am and Cities wells as shown on Figure 9a. The positions of the Jurassic salt décollement and Afton anticline east of Sublette anticline are constrained by wells located immediately north of the line of section (see Coogan and Yonkee, 1985, Fig. 7). In the kinematic model, early Eocene uplift of Sublette anticline was accomplished by late slip on a blind imbricate thrust between Sublette and Cokeville anticlines that is rooted to the Crawford trailing ramp. Two observations support this geometric interpretation: (1) Cokeville anticline plunges beneath Sublette anticline north of Cokeville in map view (Fig. 5), and (2) tight folds in the Twin Creek Limestone between Sublette and Cokeville anticlines require a corresponding amount of shortening in older strata to achieve structural balance.

The model sequence for late imbrication beneath Sublette anticline is presented in Figures 9b and c. Figure 9b shows Sublette and Cokeville anticlines after deposition of the Sublette Range Conglomerate and prior to deposition of the Wasatch Formation. Figure 9c shows the early Eocene uplift of Sublette anticline along the ramp-rooted thrust between Sublette and Cokeville anticlines. The growth of Sublette anticline above the Crawford trailing ramp is a variation of the case in which reactivated thrust slip is confined to the trailing ramp areas of thrust faults as shown in Figure 3c.

Pegram anticline. The crest of Pegram anticline was a paleotopographic boundary that formed the western margin of the early fluvial depocenter of Bear Lake Plateau basin. Subsurface data indicate that Pegram anticline is a fault-propagation fold that overlies the Crawford trailing ramp along the length of Bear Lake Plateau basin (Figs. 5 and 10) in a structural position analogous to that of Sublette anticline to the north (Figure 9a). A lateral ramp that parallels the northern margin of Bear Lake Plateau basin accounts for the westward shift in the ramp location from north to south.

Like Sublette anticline, the structural style and position of Pegram anticline resemble break-back thrust structures that form during reactivation of trailing ramp areas of thrust sheets (Figure 3c). By analogy with Sublette anticline, the paleotopographic high along the crest of Pegram anticline could have been caused by early Eocene uplift, but it is equally plausible that the high was the result of differential erosion between the resistant Nugget Sandstone on the crest of Pegram anticline and the less resistant Twin Creek Limestone on the flanks. Most of the structural relief of Pegram anticline was clearly established prior to Wasatch deposition. Successively younger Wasatch strata onlap the east limb of Pegram anticline, and the quartzite conglomerate member pinches out immediately to the east of the anticline (Fig. 10), which indicates that any minor early Eocene uplift of Pegram anticline kept pace with infilling of the early basin. Likewise, the lack of obvious angular unconformities in Wasatch strata along the flank of Pegram anticline indicates that uplift was minimal during Wasatch deposition, and the lack of extensive eastward prograding alluvial fans along the basin margin implies that early Eocene topographic relief was modest at any one time.

The position of the early basin margin above Pegram anticline was clearly controlled by the location of the underlying thrust-related structures, but the time of uplift of Pegram anticline is uncertain. The interpretation that some component of late slip beneath Pegram anticline immediately preceded Wasatch deposition is favored here by analogy to late slip on the Crawford trailing ramp immediately to the north beneath Sublette anticline.

Crawford thrust front. The Absaroka trailing ramp immediately underlies the frontal Crawford thrust plate, an interpretation that agrees with published seismic-based structure maps of the Absaroka (Dixon, 1982) and Crawford (Valenti, 1987) thrust planes. Early Eocene deformation associated with late slip on the Absaroka trailing ramp is evident from folded Wasatch strata at the surface. The Wasatch Formation dips gently (1 to 3°) west through the interior of eastern Bear Lake Plateau basin, but the dip steepens to 15 to 25° where the Wasatch has been rotated through an axial plane along the eastern basin margin (Fig. 10).

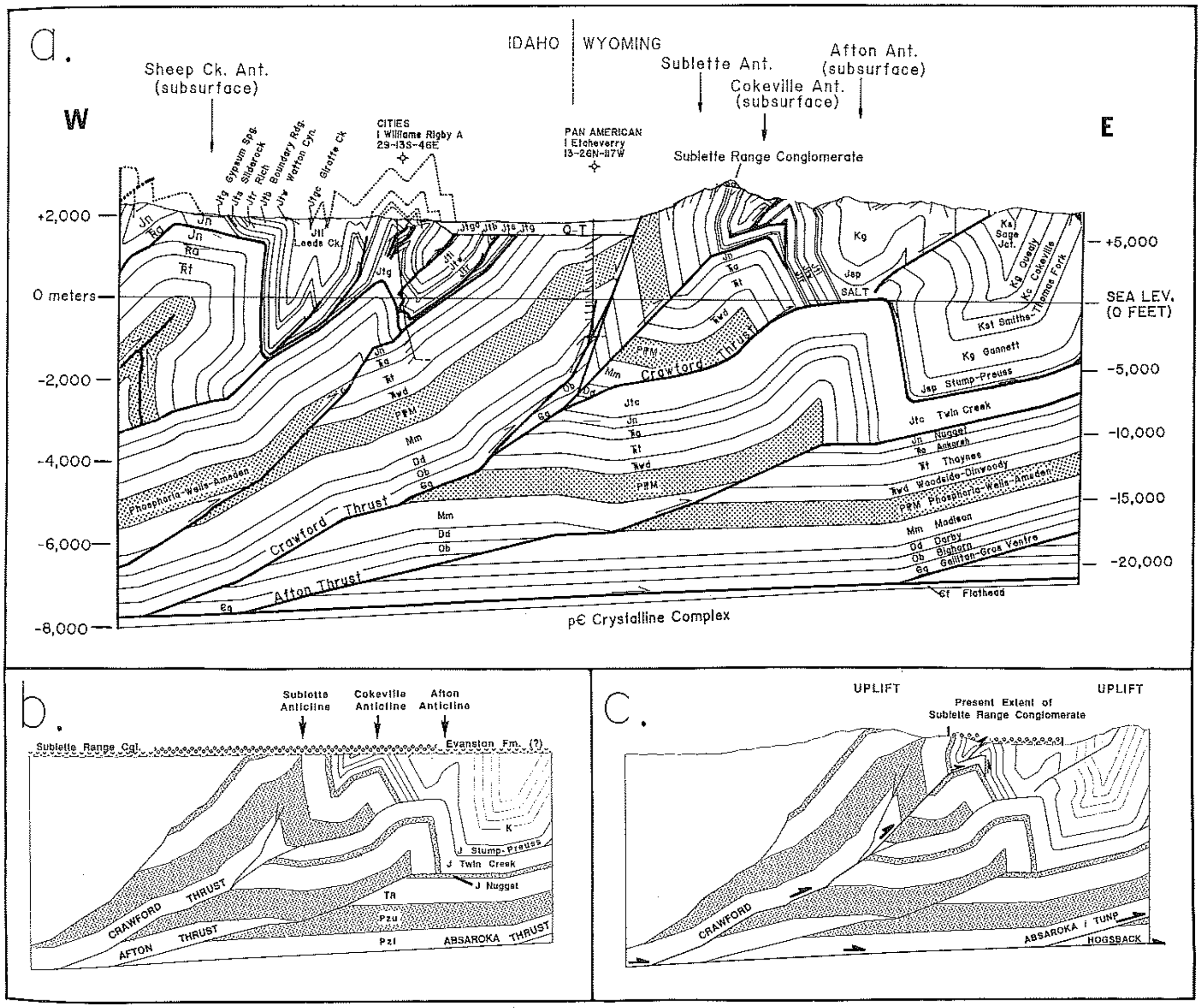


Figure 9. a, Cross section through Sublette, Cokeville, and Afton anticlines showing the present elevation and structural position of the Sublette Range Conglomerate. The trailing ramp geometry of the Crawford thrust is constrained by two wells along the west flank of Sublette anticline. Cokeville anticline plunges into the line of section from the south. Afton anticline plunges into the line of section from where it has been penetrated by two wells 9.5 km (5.1 mi) north of the section. b, Reconstruction of 9a during deposition of the Sublette Range Conglomerate. c, Reconstruction of 9a after uplift of Sublette anticline and the Sublette Range Conglomerate. Uplift was accommodated by a late ramp-rooted fault between Sublette and Cokeville anticlines immediately before or during early Eocene time. The approximate present position of the isolated outcrops of the Sublette Range Conglomerate on the east flank of Sublette anticline is shown along with the areas of uplift caused by early Eocene slip on the Crawford and Absaroka trailing ramps. Vertical scale = horizontal scale.

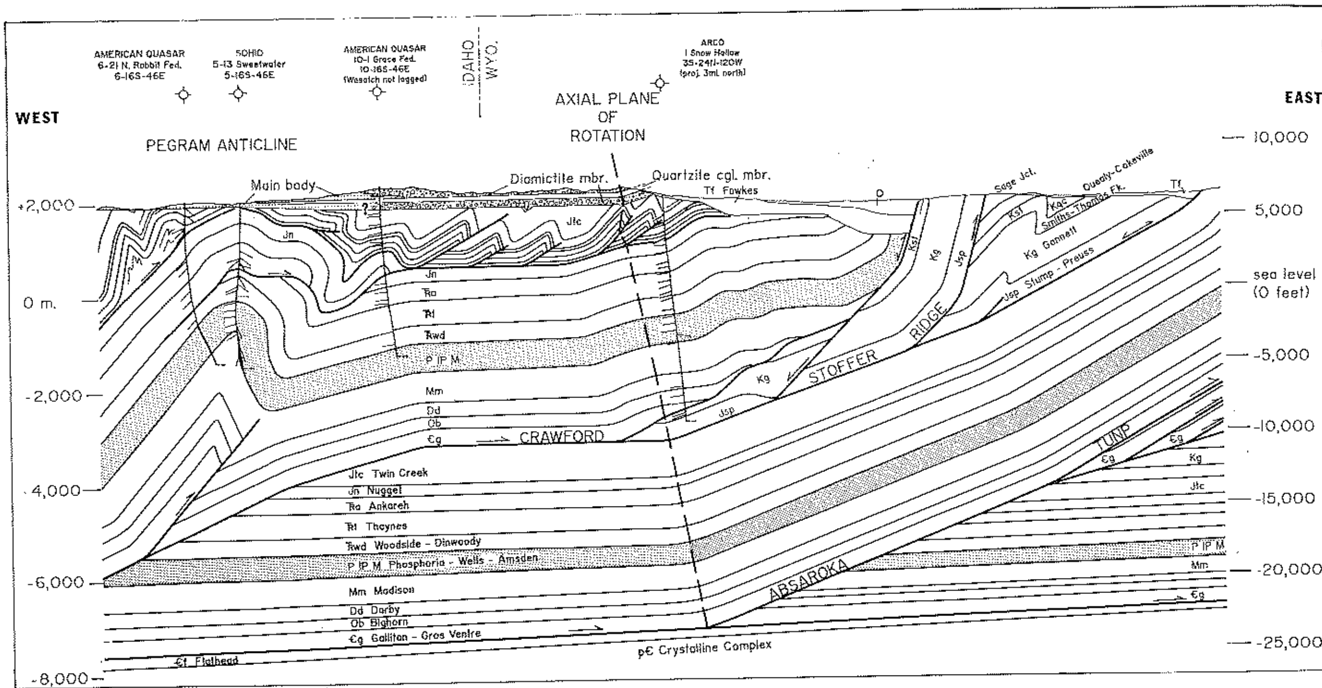


Figure 10. Cross section through eastern Bear Lake Plateau showing the geometry of the diamictite member, quartzite conglomerate member, and main body of the Wasatch Formation above Pegram anticline, Crawford thrust front, and the Absaroka trailing ramp. Pegram anticline is a fault-propagation fold above the Crawford trailing ramp that formed the western boundary of the early eastern depocenter of Bear Lake Plateau basin, as seen by the thinning of the quartzite conglomerate member along the east flank of the anticline. The leading edge of the Crawford thrust overlies the trailing ramp of the Absaroka thrust. The 20° west dip (shown by dip bars) of the Cretaceous footwall rocks at the bottom of the Arco Snow Hollow well defines the east side of an axial plane of rotation that is associated with the underlying branch point between the Absaroka thrust and the basal décollement. The Wasatch strata in the upper part of the section are also rotated approximately 20° across the axial plane. Vertical scale = horizontal scale.

The projection of the surface axial plane to depth intersects the branch point where the Absaroka thrust lifts off of the regional sole décollement in Cambrian shales at a 20° angle that is constrained by Absaroka hanging-wall dips at the base of the Arco Snow Hollow well shown in Figure 10. Deformation is clearly early Eocene in age, since the west-dipping Wasatch beds are truncated by the east-dipping middle Eocene Fowkes Formation at the surface (Figs. 5 and 10). Reactivated slip on the frontal Crawford thrust would have accomplished the same rotation of the Wasatch Formation; however, Crawford slip is precluded by the lack of substantial uplift in the Crawford trailing ramp area.

Comparison of sedimentation histories between the trailing ramp area and the leading edge of a thrust sheet is useful for determining times when through-going slip from the trailing ramp to the thrust front is possible. If thrust slip is communicated from the basal décollement, up the trailing ramp, and finally across the frontal décollement to the thrust front, a cessation of uplift in the trailing ramp region precludes any slip along the frontal length of the thrust. Sedimentation across Pegram anticline in the Crawford trailing ramp region indicates that the Crawford thrust remained locked during at least the later stages of sedimentation in Bear Lake Plateau basin. The synchronous deposition of the diamictite member along the leading edge of the Crawford thrust and the overstepping of sediments across Pegram anticline indicate that footwall uplift of the Crawford thrust front outlasted any reactivated slip and uplift on the southern Crawford trailing ramp. The duration of reactivated slip on the northern trailing ramp beneath Sublette anticline is not known. The combination of reactivated slip and uplift along the Crawford trailing ramp and footwall uplift along the Crawford thrust front represents a pattern of basin margin uplift that is also found in northern Fossil Basin and the LaBarge basin to the east.

NORTHERN FOSSIL BASIN

Structural, stratigraphic, and sedimentologic setting

Northern Fossil Basin lies within the hanging wall of the Absaroka thrust between a western basin margin adjacent to the Tunp thrust and an eastern basin margin astride the frontal trace of the Absaroka thrust (Fig. 11). Subsurface data indicate that the

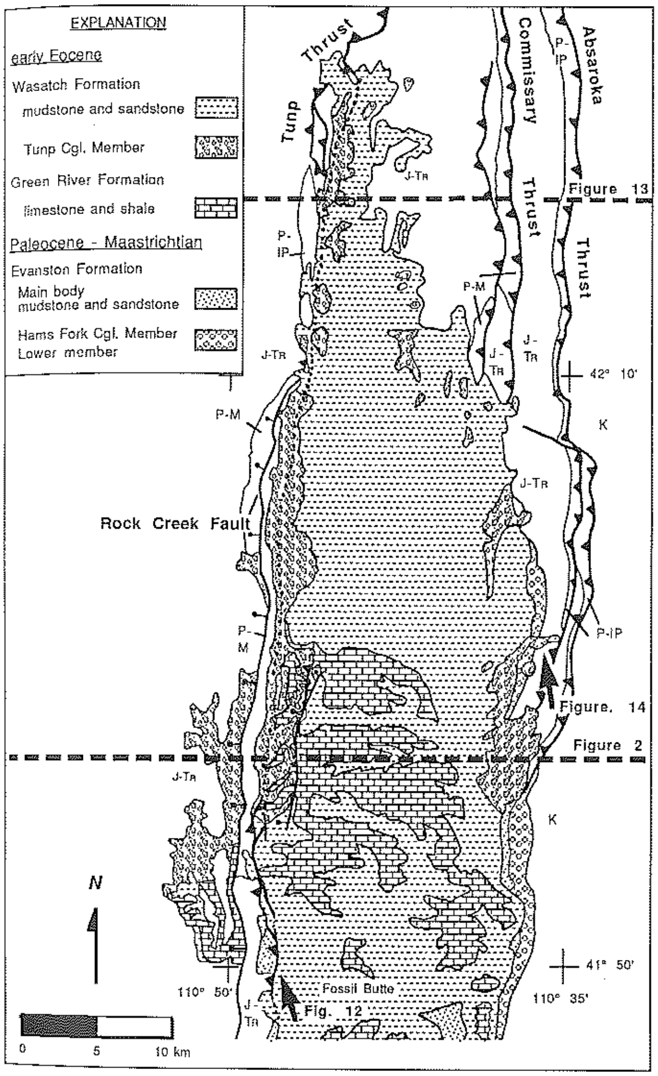


Figure 11. Generalized geologic map of latest Cretaceous and Tertiary strata in northern Fossil Basin on the Absaroka sheet. Northern Fossil basin has a symmetrical facies distribution between the coarse-grained deposits of the Tunp Member along the basin margins and the fine-grained Green River and Wasatch deposits in the basin center. The faulted western basin margin strata overlie the Tunp thrust. The trend of the eastern basin margin overlies the Hogsback thrust trailing ramp in the Absaroka footwall. P-M = Permian, Pennsylvanian, and Mississippian rocks. See Figure 5 for other stratigraphic symbols. Location of structural cross sections in Figures 2 and 13 shown by dashed lines. Location of photographs in Figures 12 and 14 shown by arrows along direction of view. From Rubey and others (1975, 1980).

Tunp thrust soles into the trailing footwall ramp of the Absaroka thrust system and that the leading edge of the Absaroka thrust plate lies above the Hogsback trailing ramp (Fig. 2).

The age and stratigraphic relationships of northern Fossil Basin (Fig. 4) were established by Oriel and Tracey (1970) and further refined by Lamerson (1982), Jacobson and Nichols (1982), and Hurst and Steidtmann (1986). Fossil Basin began as a structural and topographic depression during late Paleocene deposition of the fine-grained, fluvial main body of the Evanston Formation. Prior to that time, the Maastrichtian Hams Fork Conglomerate Member was deposited in a through-going braided stream system that crossed the area of Fossil Basin. The Hams Fork contains boulders and cobbles of Late Proterozoic quartzites that were derived from the Willard and Paris thrust sheets and transported 60 km (37 mi) across the future depocenter to their present position on the east flank of Fossil Basin (Oriel and Tracey, 1970; Salat, 1989).

Northern Fossil Basin was a well-developed north-south trending trough on the Absaroka hanging wall by early Eocene time. Both the western and eastern basin margins of northern Fossil Basin were topographically elevated at the start of Wasatch and Green River deposition, and both margins continued to be uplifted throughout infilling of the basin. The resultant early Eocene basin exhibits a symmetrical facies distribution, with the Tunp Member of the Wasatch Formation ringing the west, north, and east peripheries of the basin (Fig. 11). Hurst and Steidtmann (1986) found that the Tunp Member is dominated by matrix-supported gravels that were transported in alluvial fans toward the basin center. The Tunp Member interfingers basinward with the finer-grained fluvial lower member, main body, and Bullpen Member of the Wasatch as well as with the lacustrine limestones of the Fossil Butte and Angelo members of the Green River Formation.

Oriel and Tracey (1970) and Rubey and others (1975) considered both margins of Fossil Basin to be exclusively pre–early Eocene features. They interpreted the onlap and basinward dips of Wasatch strata along the basin margins to be the result of sedimentation onto the differentially eroded highs of the Tunp thrust hanging wall and the Absaroka thrust front, compaction of underlying sediments beneath the basin, and late block faulting. Differential erosion was important in maintaining basin margin relief throughout basin development, but it does not fully account for the early Eocene relief of the margins of Fossil Basin. For example, the lower Jurassic through Pennsylvanian ridge-forming units exposed in the Tunp hanging wall and the Absaroka thrust front are also found in subcrop beneath the center of northern Fossil Basin in the hanging wall of the Commissary thrust. As shown on Figure 2, the basal unconformity displays little relief between the resistant hanging-wall rocks and less resistant upper Jurassic footwall strata (Lamerson, 1982, Plate 6). Syndepositional uplift of erosionally resistant strata is required to explain the topographic and structural relief of both margins of northern Fossil Basin.

Structural evolution of northern Fossil Basin

Western basin margin—Tunp Thrust. The Tunp thrust has a multiphase slip history. The late Paleocene main body of the Evanston Formation (Oriel and Tracey, 1970; Lamerson, 1982; Jacobson and Nichols, 1982) overlies an angular unconformity above Jurassic and Triassic rocks of Tunp hanging wall (Rubey and others, 1975, Plate 1). Thus, initial uplift of the Tunp hanging wall occurred prior to late Paleocene time. Five features of the western margin of Fossil Basin demonstrate that the Tunp hanging wall was the site of continued uplift in the late Paleocene and early Eocene.

(1) The main body of the Evanston Formation is rotated to near vertical dips along the Tunp thrust front (Rubey and others 1975, Plate 1). The Evanston is overlain with angular unconformity by the basal conglomerate member, lower member, and main body of the Wasatch Formation (Rubey and others 1975), which indicates that uplift and east-vergent rotation occurred during latest Paleocene to earliest Eocene time.

(2) The basal conglomerate and lower members of the Wasatch Formation are also rotated to steep east dip along the southeastern edge of the Tunp hanging wall (Rubey and others, 1975, Plate 1). Intraformational unconformities between these units in the Tunp hanging wall imply that progressive tilting and uplift occurred during Wasatch deposition.

(3) The 350 m (1,150 ft) to 500 m (1,640 ft) of structural relief between equivalent late Paleocene and early Eocene stratigraphic positions in the Tunp hanging wall and footwall is adequately explained only by early Eocene uplift of the Tunp hanging wall. The unconformity between the main body of the Evanston Formation and the basal conglomerate and lower members of the Wasatch Formation lies at elevations of up to 2,200 m (7,200 ft) in the southern Tunp hanging wall (Rubey and others, 1975, Plate 1), whereas the Evanston-Wasatch contact is near 1,830 m (6,000 ft) elevation in the Tunp footwall near the basin center (Lamerson, 1982, Plate 2). Similarly, the lower member of the Wasatch Formation lies at 2,450 m (8,050 ft) elevation farther to the north in the Tunp hanging wall (Rubey and others, 1975, Plate 1), and the lower Wasatch strata are near 1,950 m (6,400 ft) elevation in the basin center (Lamerson, 1982, Plate 6). This structural relief is too large to be the result of compaction of underlying basin fill as suggested by Oriel and Tracey (1970) and must be the result of slip on the surface and buried faults along the eastern edge of the Tunp thrust hanging wall.

(4) West-dipping thrust, reverse, and normal faults along the east edge of the Tunp hanging wall truncate Wasatch strata (Figs. 11 and 12). Rubey and others (1975, p. 14) interpreted these faults as high-angle block faults related to the post-thrust regional extension seen in the Basin and Range Province to the west. The west-dipping Rock Creek fault (Fig. 11) is a late listric normal fault that is shown by well control to sole into the Tunp thrust at depth (Lamerson, 1982, Plate 2). West-dipping faults that cut Wasatch strata east of Rock Creek exhibit changes from reverse to normal offset along the strike of the same fault trace (Fig. 11; Rubey and others, 1975, Plate 1). These west-dipping faults are probably imbricates to the Tunp thrust that were locally reactivated during the later episode of normal slip on the Tunp thrust plane that was documented for the Rock Creek fault. To the north, these west-dipping faults are overlain by the Tunp Member of the Wasatch Formation, which indicates that they initiated in the early Eocene prior to deposition of the overlying strata (Rubey and others, 1975, p. 14) and prior to the inception of regional extension in late Cenozoic time (Stewart, 1978).

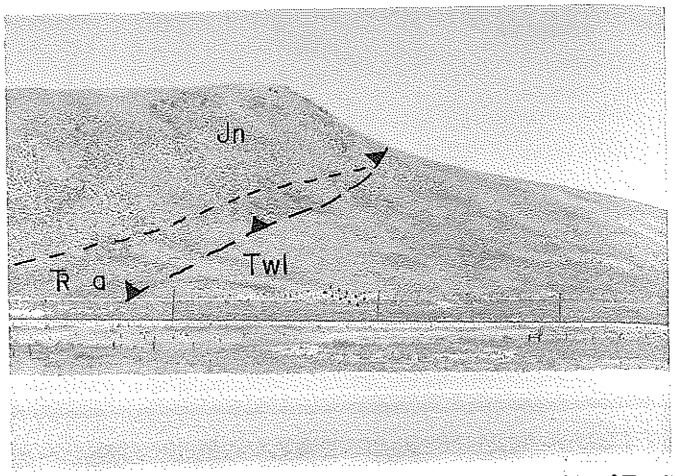


Figure 12. View to the north of a thrust fault on the west side of Fossil Basin that places the Triassic Ankareh Formation (Ꝛa) and the Jurassic Nugget Sandstone (Jn) over the lower member of the Wasatch Formation (Twl). Location is shown on Figure 11.

(5) Thin limestones of the Fossil Butte Member of the Green River Formation (Fig. 4) dip up to 58° east where they are interbedded with the Tunp Member of the Wasatch Formation above the buried Tunp thrust and its imbricates (Rubey and others 1975, Plate 1; 1980, Plate 1). These obvious tectonic dips demonstrate that uplift of the Tunp hanging wall continued into or beyond the later stages of basin deposition. The concealment of the thrust beneath the Tunp Member seems to have been one reason why earlier workers did not recognize early Eocene slip on the fault. Burial of the fault should be an expected feature along the western basin margin where sequential uplift of resistant hanging wall rocks provided elevated source areas for alluvial fans that continually buried the fault trace during sediment transport into the basin.

Figure 13 is a detailed cross section through the northern tip of Fossil Basin in the area of shallowest basin fill that illustrates the relationship between the Tunp thrust, the Absaroka thrust sheet, and the Tunp Member of the Wasatch Formation. The branching relationship between the Tunp thrust and the Absaroka trailing ramp is geometrically constrained by the intersection of the steep west-dipping panel of the Tunp hanging wall with the seismically defined position of the Absaroka trailing ramp (Dixon, 1982). The subsurface geometry of the Tunp thrust footwall is projected from the Amerada-Sunmark #1-19 Federal well (Sec. 19, T27N, R117W) 6.5 km (4 mi) north of the section and the Amerada #1-23 Federal well (Sec. 23, T25N, R118W) 14.5 km (9 mi) south of the section. Both wells penetrated the Tunp thrust, the Commissary thrust, and the underlying duplex of Cambrian and Ordovician strata. The duplex system in the Tunp footwall did not contribute to early Eocene uplift of the western margin of Fossil Basin. The Commissary and Absaroka thrusts

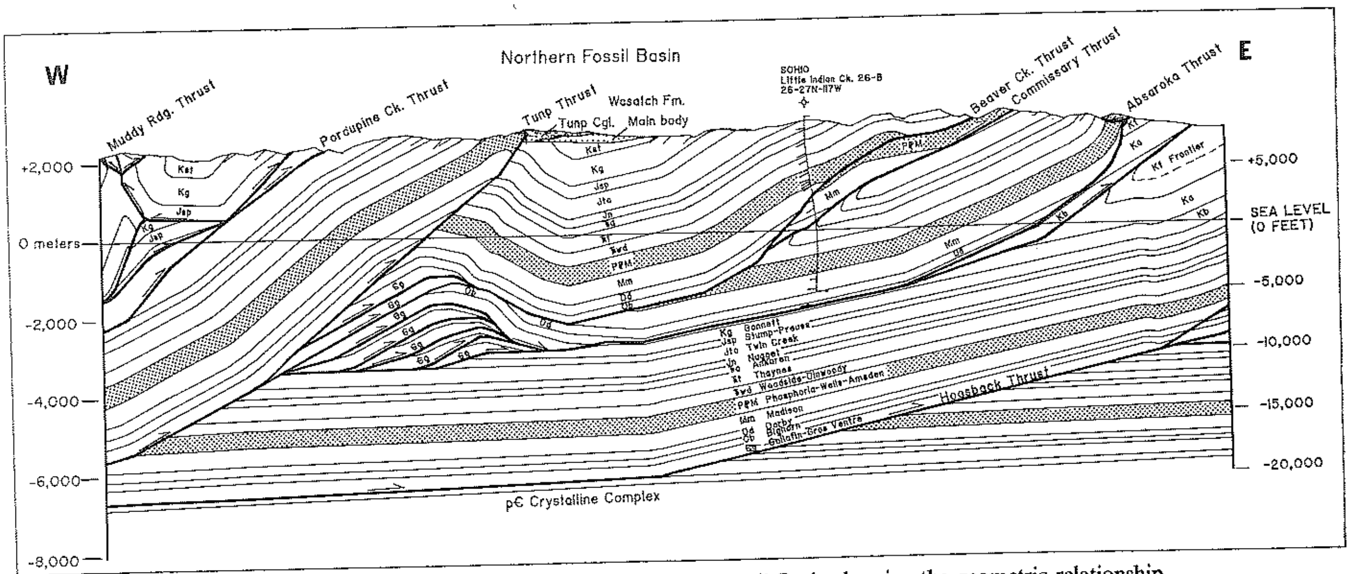


Figure 13. Cross section through the northern end of Fossil Basin showing the geometric relationship between the Tunp Member of the Wasatch Formation and its source area on the Tunp thrust hanging wall. Notice that the Tunp thrust soles into the Absaroka trailing ramp and that the eastern basin margin overlies the Hogsback trailing ramp. Vertical scale = horizontal scale.

are roof thrusts to the duplex system, and both thrusts are overlain by late Paleocene strata along strike (Figure 2; Lamerson, 1982, Plates 2 and 6), indicating that slip along the duplex imbricates predates early Eocene uplift. Consequently, reactivated slip along the Absaroka trailing ramp bypassed the mechanically locked Absaroka footwall flat. Slip on the Tunp thrust is the only possible mechanism for the early Eocene uplift along the western margin of northern Fossil Basin.

Eastern basin margin—Absaroka thrust and Hogsback ramp. The leading edge of the Absaroka thrust sheet exhibits crosscutting and overlapping relationships that firmly establish the age of slip along the Absaroka thrust itself as well as the age of footwall uplift above the Hogsback trailing ramp. The last, minor reactivated slip on the frontal Absaroka thrust during the late Maastrichtian or early Paleocene cut the Hams Fork Conglomerate, and the leading edge of the Absaroka thrust sheet was onlapped and overlain by the late Paleocene main body of the Evanston (Lamerson, 1982, Plates 2 and 9). The pre–late Paleocene uplift was the earliest expression of the eastern basin margin along the leading edge of the Absaroka thrust sheet.

The leading edge of the Absaroka thrust sheet was the site of uplift throughout early Eocene deposition of the Wasatch and Green River formations. The late Paleocene overlap of the Absaroka fault precluces reactivated slip as a mechanism for uplift, yet the main body of the Evanston Formation is rotated to 15° west dip along the eastern basin margin (Rubey and others, 1975), and it is successively overlain by the main body and Tunp Member of the Wasatch Formation (Fig. 14). Lamerson (1982, p. 298–299) first proposed that slip along the underlying Hogsback thrust ramp was responsible for uplift of the Absaroka footwall, the mechanically locked Absaroka thrust, and the overlying hanging-wall and overlap strata. Hurst and Steidtmann (1986) concluded that the alluvial fans of the Tunp Member of the Wasatch along the east side of northern Fossil Basin were shed from this area of uplift.

The relief on the eastern margin of Fossil Basin changes drastically north and south of the area shown in Figure 2 (Fig. 1). The relief is controlled by the position and amount of structural relief of the Hogsback trailing ramp beneath the Absaroka thrust sheet as well as by the erosional resistance of strata carried in the Absaroka hanging wall. There is direct correspondence between the position of the eastern basin margin and the underlying Hogsback trailing ramp. The westward deflection in the position of the Hogsback ramp from that shown in Figure 2 to that shown in Figure 13 parallels and underlies the westward deflection of the eastern margin of northern Fossil Basin (see also Dixon, 1982, Fig. 18). At the latitude of the area shown in Figure 13, footwall uplift above the Hogsback trailing ramp elevated most of the Absaroka footwall flat and was not simply limited to the leading edge of the Absaroka thrust plate. South of the area shown in Figure 2, the Hogsback trailing ramp lies immediately beneath the outcrop and subcrop trace of the Absaroka thrust (Lamerson, 1982, Plates 8, 5, 4, and 10), and the relief of the eastern margin of Fossil basin is correspondingly low. Between the area shown in Figure 2 and the Little Muddy Creek area (Fig. 1), resistant lower Jurassic through Pennsylvania rocks form the subcrop toe of the Absaroka thrust (Lamerson, 1982, Plate 8). In this case, the low basin margin relief cannot be attributed to differential erosion and must be the result of the amount of footwall uplift that was imparted by slip on the underlying Hogsback ramp. South of

Figure 14. View to the north of the eastern margin of Fossil Basin at Trail Creek. Foreground and small peak in background are capped by the flat-lying Tunp Member of the Wasatch Formation (Twt), which overlies steeply dipping Triassic (TR) and Jurassic (J) rocks of the Absaroka hanging wall as well as the steeply dipping Hams Fork Member (Keh) and lower member (Kel) of the Evanston Formation. Onlap by the shallowly west dipping lower member (Twl) and main body of the Wasatch Formation (Tw) is seen to the left. The progressive tilting and erosion of latest Cretaceous through early Eocene strata were the result of Paleocene through early Eocene footwall uplift above the Hogsback trailing ramp.

Little Muddy Creek, the low relief of the basin margin is enhanced by the erosionally soft upper Jurassic and lower Cretaceous rocks carried at the toe of the Absaroka thrust (Lamerson, 1982, Plates 5, 4, and 10), but the fact that the Absaroka thrust trace remains buried beneath late Paleocene and early Eocene basin sediments indicates that early Eocene footwall uplift was minimal along the southern length of the thrust. Finally, the early Eocene eastern margin of Fossil Basin disappears south of the Utah-Wyoming boundary where the Hogsback trailing ramp lies east of the buried Absaroka thrust trace and early Eocene deposits in southeastern Fossil Basin are continuous with those in the southwestern Green River Basin (Fig. 1).

Northern Fossil Basin provides the clearest evidence for the combined effect of reactivated thrust uplift along the trailing basin margin and footwall uplift along the leading basin margin. Unlike Bear Lake Plateau basin, Fossil Basin exhibits a symmetrical filling about a fixed depocenter, which implies that reactivated thrust uplift and erosion above the Absaroka trailing ramp kept pace with footwall uplift and erosion of the Absaroka thrust front above the Hogsback trailing ramp.

LABARGE BASIN

Structural, stratigraphic, and sedimentologic setting

A small earliest Eocene piggyback basin, which is informally referred to as the LaBarge basin, is located between the trailing ramp and the leading edge of the Hogsback thrust sheet west of LaBarge, Wyoming (Fig. 1). Later early Eocene strata that overlap the piggyback basin above an angular unconformity define the western edge of the greater Green River Basin. Figure 15 shows the structural position of the Hoback, Wasatch, and Green River formations relative to the trailing ramp, leading edge, and footwall imbricates of the Hogsback thrust at higher structural levels and to the Moxa Arch at basement level. Seismic data indicate that the Hogsback trailing ramp dips approximately 13° west from its position immediately west of LaBarge basin (Fig. 15) to its position beneath the Absaroka thrust plate (Fig. 13). The Meridian and Fort Hill thrusts are hanging-wall imbricates of the Hogsback thrust that sole into the area of transition between the trailing footwall ramp and the frontal footwall décollement of the Hogsback thrust (Fig. 15). The buried traces of the Meridian and Fort Hill thrusts lie on the western margin of LaBarge basin (Fig. 15). The LaBarge and Calpet thrusts are conjugate footwall imbricates of the Hogsback thrust that are linked to the Hogsback trailing ramp along a Cretaceous décollement in the Hogsback footwall (Fig. 15). The LaBarge and Calpet thrusts intersect in a triangle zone (Teal, 1983) east of the Hogsback thrust front where the displacement on the east-dipping Calpet thrust was transferred to the upper part of the west-dipping LaBarge thrust in the subsurface.

The age and intertonguing stratigraphic relationships between Paleocene through early Eocene strata of the LaBarge basin were first treated by Oriel (1962, 1969) and refined by Dorr and Gingrich (1980) as presented on Figure 4. The Chappo Member of the Wasatch Formation is the oldest Cenozoic unit exposed at LaBarge. The type locality of the Chappo Member (Oriel, 1962) lies east of the Hogsback thrust and contains late Paleocene (Tiffinian) strata (Dorr and Gingrich, 1980). Well correlations indicate that conglomerates at the base of the late Paleocene Chappo Member are correlative to conglomerates in the Hoback Formation east of the Hogsback thrust front (Oriel, 1969; Dorr and Gingrich, 1980). Earliest Eocene (Clarkforkian) strata were also mapped as the Chappo Member by Oriel (1969) in the hanging wall of the Hogsback thrust. The presence of two distinct outcrop belts of the Chappo Member with different ages and structural affinities is a source of confusion, and a redefinition of the Chappo Member should be considered in the future. In this study, the two stratigraphic units are discussed separately as the late Paleocene Chappo Member and the early Eocene Chappo Member, and the late Paleocene Chappo Member is grouped with the Hoback Formation in the subsurface in Figure 15.

The early Eocene LaBarge and Lookout Mountain Conglomerate members of the Wasatch Formation overlie both outcrop belts of the Chappo Member with angular unconformity. The Lookout Mountain Conglomerate consists of interbedded diamictite, conglomerate, sandstone, and mudstone that resemble the Tunp Member of Fossil Basin and the diamictite member of Bear Lake Plateau basin. The diamictites of the Lookout Mountain Conglomerate contain angular clasts of Mesozoic and Paleozoic strata derived from the trailing ramp area of the Hogsback thrust plate, including clasts up to 2 m (6 ft) in length (Oriel,

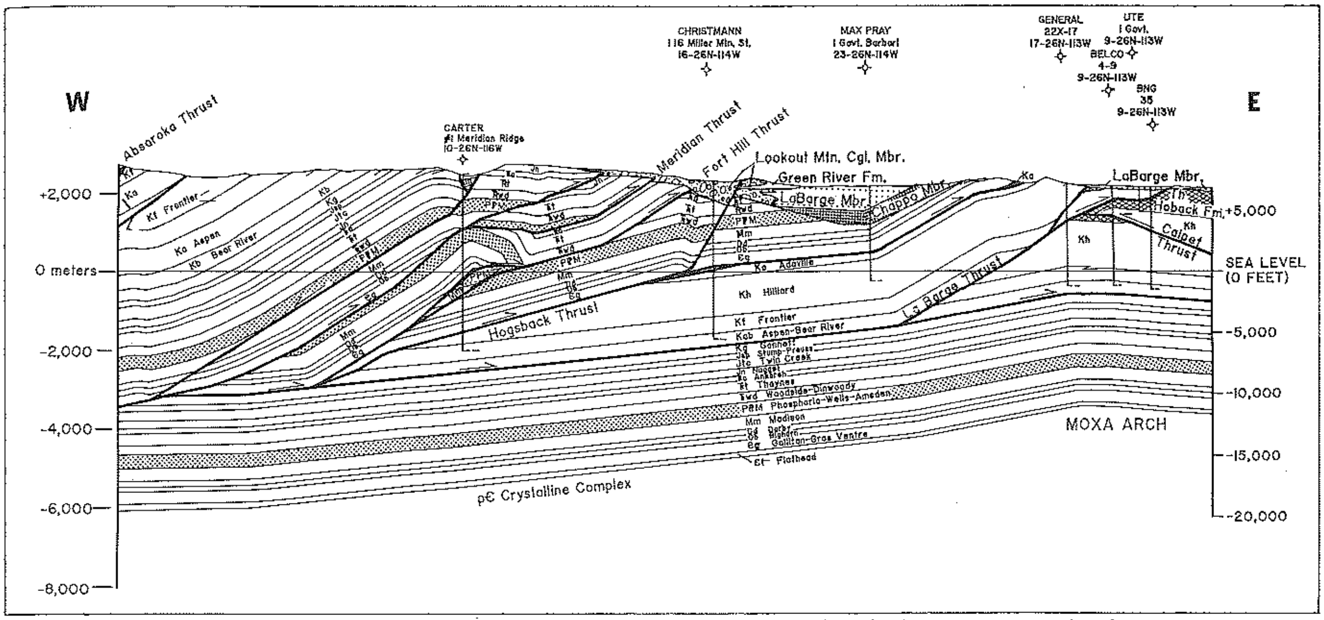


Figure 15. Cross section through the Hogsback thrust and the Moxa Arch in the LaBarge area, showing the geometry of the folded Chappo Member of the Wasatch Formation in LaBarge basin. LaBarge basin lies between an area of uplift above the Hogsback trailing ramp and the Hogsback thrust front above the LaBarge thrust ramp. The LaBarge and Calpet thrusts are conjugate footwall imbricates to the Hogsback thrust that form a triangle zone above the Moxa Arch. Later deposition along the western margin of the Green River Basin overlapped both LaBarge basin and the triangle zone with angular unconformity. The late western margin of the Green River Basin is defined by the Lookout Mountain Conglomerate alluvial fan deposits that lie above the Meridian thrust. Vertical scale = horizontal scale.

1962). These gravels are commonly matrix supported and were transported eastward on alluvial fans. The Lookout Mountain Conglomerate interfingers eastward with the fine-grained fluvial deposits of the LaBarge Member, New Fork Tongue, and upper tongue of the Wasatch Formation as well as with the lacustrine limestones of the Fontenelle tongue and the middle tongue of the Green River Formation.

Structural evolution of the LaBarge basin

LaBarge basin. LaBarge basin lies immediately west of Hogsback thrust front, the LaBarge and Calpet footwall imbricates, and the Moxa Arch basement uplift (Fig. 15). The Moxa Arch was uplifted by late Campanian time and predates all other structures in the area (Wach, 1977). The Hogsback thrust was fully emplaced by the late Paleocene (Warner and Royse, 1987). The thrust cuts the late Paleocene Hoback Formation 5 km (3 mi) north of the area shown in Figure 15 (Blackstone, 1979, Plate 6; Warner and Royse, 1987) and is in turn overlain by the earliest Eocene part of the Chappo Member (Dorr and Gingrich, 1980).

LaBarge basin developed after some 4.5 km (2.7 mi) of Cretaceous through Paleozoic strata were eroded from the Hogsback hanging wall. The early Eocene Chappo Member of the Wasatch Formation is the oldest unit exposed above the eroded Hogsback thrust sheet (Dorr and Gingrich, 1980). The early Eocene Chappo Member coarsens westward above the Hogsback thrust and contains clasts derived from Mesozoic and Paleozoic units that are exposed above the Hogsback trailing ramp to the west (Oriel, 1962). The westward onlap of Wasatch strata is evidence that considerable topographic relief existed above the Hogsback ramp in the area of known early Eocene uplift that was discussed earlier for eastern Fossil Basin. The onlap of the early Eocene Chappo Member along the eastern margin of LaBarge basin is more subtle, but the geometry of the Hogsback thrust sheet requires that deposition in the piggyback basin followed initial footwall uplift of the Hogsback thrust front.

Paleozoic rocks along the leading edge of the Hogsback thrust front did not form a high topographic boundary along the eastern basin margin prior to slip on the underlying LaBarge thrust. By restoring the Paleozoic strata to the 6° west dip that they held prior to LaBarge thrust slip, the unconformity beneath the early Eocene Chappo becomes essentially flat between its location above Mississippian rocks in the Max Pray well and its location above Cambrian rocks along the Hogsback thrust front (Fig. 15). The eastward onlap of the early Eocene Chappo Member onto the unconformity surface (Dorr and Gingrich, 1980) implies that an increment of uplift preceded Chappo deposition. The present 10 to 25° west and southwest dips of the early Eocene Chappo Member along the eastern basin margin demonstrate that uplift continued after Chappo deposition. Uplift of the eastern basin margin was caused by slip along the LaBarge thrust ramp in the footwall of the Hogsback thrust. This uplift occurred prior to deposition of the flat-lying LaBarge Member, which overlies the early Eocene Chappo Member with angular unconformity at the surface (Oriel, 1969).

Earliest Eocene slip along the LaBarge and Calpet thrusts had to be balanced by coeval slip along the Hogsback trailing ramp because the thrusts are geometrically and mechanically linked along the Cretaceous décollement in the Hogsback footwall (Fig. 15). Thus, footwall uplift of the eastern basin margin during early Eocene Chappo deposition was matched by late slip and uplift along the Hogsback trailing ramp on the western basin margin.

Western margin of Green River Basin. The present western margin of the Green River Basin developed with the onset of LaBarge Member deposition following uplift and erosion of the underlying piggyback basin strata and the LaBarge and Calpet thrust hanging walls. The depth of the contact between the LaBarge and Chappo members encountered in the Max Pray well (Fig. 15) indicates that the LaBarge Member initially infilled the piggyback basin above the Hogsback thrust sheet while onlapping the Chappo Member to the east.

Above the LaBarge Member, the western Green River Basin margin is defined by an eastward progression from alluvial fan deposits of the Lookout Mountain Conglomerate through fine-grained fluvial strata of the Wasatch Formation and into lacustrine limestones of the Green River Formation. Lookout Mountain Conglomerate clasts were clearly generated from the area of the Hogsback thrust sheet that overlies the Hogsback trailing ramp. Late slip on the Hogsback trailing ramp likely caused minor late uplift of the western basin margin based on surface and subsurface observations. Gentle warping of the LaBarge Member above the LaBarge thrust mapped by Oriel (1969) demonstrates that some minor slip reached the thrust front during Lookout Mountain deposition. However, most of the reactivated slip along the Hogsback trailing ramp probably bypassed the leading edge of the steeply dipping and mechanically locked footwall imbricates. Instead, some late shortening could have been accommodated along the Meridian thrust or its imbricates (Fig. 15). The Meridian thrust soles into the Hogsback trailing ramp area in a structural style and position analogous to that of the Tunp thrust above the Absaroka trailing ramp. Surface observations that support early Eocene slip on the Hogsback trailing ramp include the following: (1) Down-to-the-east faults mapped as block faults by Oriel (1969) both cut and overlie the Lookout Mountain Conglomerate above the Meridian thrust in a setting analogous to the Tunp Member and the Tunp thrust of western Fossil Basin, (2) gentle folds in the Lookout Mountain Conglomerate overlie the subcrop of Meridian thrust imbricates, and (3) Wasatch diamictites reported by Tracey and others (1961) were directly involved in thrusting 27 km (16 mi) north of the LaBarge area at the north end of the Hogsback thrust plate. Uplift of the eastern margin of Fossil Basin above the Hogsback trailing ramp during this period also implies continued late uplift along the western margin of the LaBarge area. In addition, Delphia and Bombolakis (1988) interpreted break-back thrust imbrication above the Hogsback trailing ramp based on palinspastic reconstructions of the Hogsback thrust sheet 50 km (30 mi) south of the LaBarge area.

DISCUSSION: CAUSE OF REGIONAL THRUST REACTIVATION

Beginning in latest Paleocene, the eastern half of the Wyoming-Idaho-Utah thrust belt experienced a fundamental change in structural style and mechanical behavior that culminated in the development of early Eocene piggyback basins. Reactivated slip along the trailing ramps of the Crawford and Absaroka thrusts and late slip along the Hogsback trailing ramp marked an episode of break-back thrusting that contrasts with the earlier main-phase, foreland-directed thrust sequence. The initiation of break-back thrusting during the early Eocene raises the question of whether the anomalous pattern of thrust slip was a response to unique mechanical conditions in the thrust belt during the final phase of regional shortening. A likely reason for break-back thrust episode is the impedance of further slip on the frontal Hogsback-Darby-Prospect thrust zone, coupled with the accommodation of further shortening by reactivation of mechanically interconnected thrusts behind the thrust front. The mechanical interconnection of the Crawford, Absaroka, and frontal thrusts is outlined below.

Mechanical organization of thrusts

Thrust faults in the Wyoming-Idaho-Utah thrust belt are divided into two separate geometric and mechanical thrust systems from west to east (Coogan, 1987). The separate thrust systems are defined by thrust surfaces that branch from a common basal décollement. The Willard, Paris, and Meade thrusts make up a western thrust system that is defined by a master sole décollement at the base of the thick Late Proterozoic clastic sequence of the Cordilleran miogeocline. In contrast, the Crawford, Absaroka, Hogsback-Darby, and Prospect thrusts are linked along a master sole décollement in Cambrian shales at the base of the thin Wyoming shelf sequence (Fig. 2).

The geometric, temporal, and mechanical separation between these two systems is displayed in the Wasatch Mountains near Ogden, Utah. There, the Willard thrust is folded above a basement anticlinorium associated with the underlying eastern thrust system (Royse and others, 1975, Plate II). The Archean Farmington Canyon Complex forms the core of a mid-crustal trailing ramp anticline in the interior of the eastern thrust system, where the basement rocks were translated onto a décollement in the Cambrian Gros Ventre shale of the Wyoming shelf sequence. The uplift onto and translation across the Cambrian décollement represent basement shortening that is equivalent to thin-skinned shortening along the successively eastward-younging Crawford, Absaroka, and Hogsback thrusts (Coogan, 1987). The successive uplift associated with Crawford, Absaroka, and Hogsback thrusting folded the Willard thrust and mechanically isolated the western thrust system from further slip. The uplift of the western thrust system above the interior ramp of the eastern thrusts is the clearest example of the interior ramp-supported uplift mechanism discussed by Schmitt and Steidtmann (1990).

The basement anticlinorium of the Wasatch Mountains was progressively exposed as a clastic source area during main-phase Crawford (DeCelles, 1988) and Absaroka (Schmitt and Steidtmann, 1990) thrust slip, and it was the source area for conglomerates of the Wasatch Formation during the last phase of regional shortening (Crawford, 1979). The final early Eocene interior uplift of the Wasatch Mountains during deposition of the Wasatch Formation was matched to the east by uplift of piggyback basin margins above the Crawford, Absaroka, and Hogsback thrusts. Early Eocene slip along the trailing ramps of these thrusts is one expression of their mechanical interconnection along the Gros Ventre sole décollement of the eastern thrust belt. Break-back thrust slip along the Crawford and Absaroka trailing ramps accommodated shortening along the Gros Ventre décollement system during the cessation of slip on the frontal Hogsback thrust.

Causes of slip impedance along the thrust front

The cause of frontal slip impedance during the final stages of shortening in a thrust belt is part of a larger problem of why thrust belts stop moving. Recent material wedge models of thrust belts have been successful in explaining how thrust belts propagate (Chapple, 1978; Davis and others, 1983). Given sufficient push at the rear of the belt, a thrust belt with reasonable material properties should continue to maintain the shape required for it to move indefinitely (Chapple, 1978, p. 1192). Final Paleocene through early Eocene shortening in the Wyoming-Idaho-Utah thrust belt geographically and temporally overlapped shortening of basement and cover rocks in the Wyoming foreland (Perry and Schmidt, 1988). Hamilton (1988) correlates the end of foreland thin-skinned and basement shortening with a slowing of plate convergence rates in early Paleogene time. On a regional scale, the cessation of thrusting is probably the result of a diminished tectonic push at the rear of the thrust belt. On a smaller scale, the impedance of slip on the frontal thrusts during the last phase of thrust shortening can be related to mechanical interaction of the thrust front with specific and general features of the foreland. Four mechanisms for the impedance of slip along the thrust front are examined as explanations for reactivation on trailing thrusts: (1) buttressing against foreland basement uplifts, (2) thrust deflection above foreland basement warps, (3) decreased taper of the regional thrust wedge caused by foreland subsidence and thrust belt erosion, and (4) stratigraphic and rheologic changes across the eastward-thinning sedimentary section from thrust belt to foreland. These mechanisms probably interacted in different locations and at different scales during the final episode of thrust shortening.

Foreland buttressing. The term *foreland buttress* is limited in this chapter to cases in which frontal thin-skinned thrust sheets abut preexisting or coeval foreland basement uplifts. The buttress geometries consist of interconnected and crosscutting networks of west-dipping thin-skinned thrusts and east-dipping basement-rooted thrusts. The best example of a buttressed thrust front is located 100 km (62 mi) north of the LaBarge basin. There, west-dipping frontal thrusts of the Prospect thrust zone intersect coeval, east-dipping, basement-involved thrusts of the Gros Ventre uplift. Royse (1985) and Hunter (1988) demonstrated that the Prospect thrust system is composed of a break-back thrust sequence in the area where the frontal Granite Creek thrust was overridden by the basement-rooted Cache Creek blind thrust. Successive hanging-wall imbricates (Prospect, Little Granite Creek, Bull Creek, Game Creek, and Bear thrusts) to the Granite Creek thrust formed in a westward-younging sequence (Hunter, 1988). Although the northern Prospect thrust front provides a well-documented type example of foreland buttressing in the Wyoming thrust belt, the Gros Ventre buttress is too distant to have influenced Hogsback thrust slip to the south.

A foreland buttress model for the Moxa Arch was advanced by Kraig and others (1988) along the southern length of the Prospect thrust north of LaBarge, Wyoming. Dixon (1982), Royse (1985), and Kraig and others (1988) demonstrated that the Moxa Arch rapidly increases in structural relief north of the area illustrated in Figure 15. Thirty km (19 mi) to the north, the seismic expression of the Moxa Arch shows approximately 1,850 m (6,070 ft) of structural relief on the basement surface that was uplifted along a reverse fault on the west side of the arch (Kraig and others, 1988). Kraig and others (1988) interpret that pre-Paleocene slip on the basement fault propagated upward to a Triassic décollement in the cover rocks that was shared by an early frontal segment of the Prospect thrust. The pre-Paleocene basement uplift and the shared slip planes for thrust belt and foreland structures imply that the Moxa Arch was a preexisting buttress that guided slip on the frontal Prospect thrust in late Paleocene and early Eocene. It is plausible that foreland buttressing of the southern Prospect thrust could have influenced final slip on the northern Hogsback thrust immediately to the south near LaBarge (Fig. 1). At LaBarge, however, the Moxa Arch is a gentle, unfaulted basement warp where the local influence of the arch on the geometry and kinematics of the frontal thrusts is not obvious.

Foreland basement warp. Wiltschko and Eastman (1983) experimentally demonstrated that foreland basement warps can concentrate stress and deflect fault trajectories in the overlying sedimentary section during horizontal compression. Their photoelastic model of the Moxa Arch (Fig. 16) illustrates the stress distribution along a block of photoelastic material that is loaded from the left and fixed on the right above a rigid substrate (Wiltschko and Eastman, 1983, Fig. 4b). The highest stresses are concentrated along the top of the block on the fixed side of the arch. Wiltschko and Eastman (1983) indicate that shallow faults should form through this area, which generally agrees with the

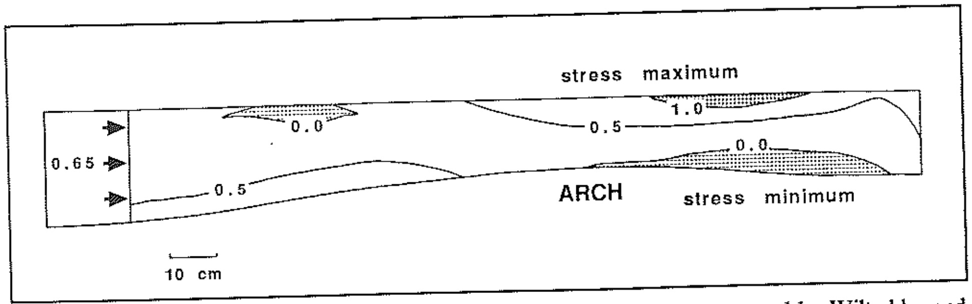


Figure 16. Stress distribution in photoelastic model of the Moxa Arch constructed by Wiltschko and Eastman (1983). A wedge of high polymer plastic was loaded laterally from the left and fixed on the right. Stress is represented by fringe orders that result from a 0.65 order load. The area of highest stress (0.5–>1.0 order) is located above the arch crest along the top of the wedge, indicating that shallow faults should form through this area. The area of lowest stress (0 order) lies in a stress shadow localized at the base of the wedge along the right flank of the arch. The model implies that foreland basement warps serve to deflect deep faults to shallower structural levels and that deep thrusts should not propagate beyond the crest of basement warps. From Wiltschko and Eastman (1983, Figure 4b).

observed position of the Calpet and LaBarge thrusts at shallow structural levels as shown in Figure 15. In contrast, the lowest stresses occur along the base of the block to the right of the arch crest. This stress shadow would impede thrust propagation near the basement-cover contact on the foreland side of the arch. Wiltschko and Eastman (1983) concluded that the Moxa Arch primarily served to deflect deep faults west of the arch to shallower structural levels above the arch and that deep thrusts should not propagate east of the arch. The model reveals that gentle basement warps can provide a tip, or sticking point, to further slip along the basal décollement near the basement cover interface in front of a thrust belt.

Foreland buttressing and thrust deflection above basement warps cannot be the only mechanisms responsible for slip impedance along the frontal thrust belt and reactivated slip in the interior. For example, Paleocene and post–early Eocene reactivated slip is documented far to the south of Moxa-Hogsback convergence for the Medicine Butte thrust along the western margin of southern Fossil Basin (Lamerson, 1982). Regional scale mechanisms for frontal slip impedance are required to explain late thrusting in the interior of the belt in areas far from foreland buttresses and basement warps.

Thrust belt taper. Late thrusting in the interior of thrust belts has been cited as a mechanism for maintaining the appropriate shape for stable sliding of thrust belts above a regional basal décollement (Woodward, 1987). Davis and others (1983) proposed that thrust belts propagate as brittlely deforming wedges above weak, through-going décollements as illustrated in Figure 17a. When subjected to a lateral load, such wedges deform internally to a critical taper, at and beyond which they slide stably. Davis and others (1983) related the topographic slope of the wedge α and the basal décollement slope β (Figs. 17a and 18) to the material properties of the wedge and décollement. The line defining the range of stable wedge shapes on Figure 18 divides potential wedge shapes into a supercritical field for stably sliding wedges with greater than critical taper and a subcritical field for wedges that experience internal deformation at less than critical taper.

Woodward (1987) proposed that late faulting in the interior of thrust belts is one expected feature of subcritical wedges. Initial shortening of a subcritical thrust belt could be accommodated by internal thrusts that elevate the rear of the thrust belt and reestablish the critical taper for stable sliding to the front of the belt. Woodward (1987) implied that break-back thrusting should be a common process throughout thrust belt evolution for critical wedge models to work. Such internal shortening and uplift would counter the taper-reducing uplift caused by the successive emplacement of thrust sheets at the front of the wedge (Woodward, 1987). The fact that break-back thrusting is only recognized for the final phase of shortening in the Wyoming-Idaho-Utah thrust belt implies that critical or supercritical taper was maintained prior to that time. Uplift along the interior ramp beneath the Wasatch basement is one mechanism that was overlooked by Woodward (1987) for maintaining an elevated rear of this thrust belt during successive emplacement of the Crawford, Absaroka, and Hogsback thrusts at the front of the belt. However, early Eocene uplift in the Wasatch Mountains area should have also contributed to maintaining wedge taper. Thus, a taper-reducing mechanism that is unique to the final phase of shortening is needed to explain the late break-back thrust events in a critical wedge model.

Differential subsidence of the Wyoming foreland is one taper-modifying mechanism that is unique to the period of final shortening in the thrust belt. Paleocene and Eocene subsidence in the foreland was the result of topographic and tectonic loads from main-phase emplacement of basement uplifts as well as sediment loads from the detritus that was shed from these uplifts into the surrounding intermontane basins. Shuster and Steidtmann (1988) demonstrated that the northern Green River Basin experienced a period of rapid tectonic subsidence in the Paleocene and early

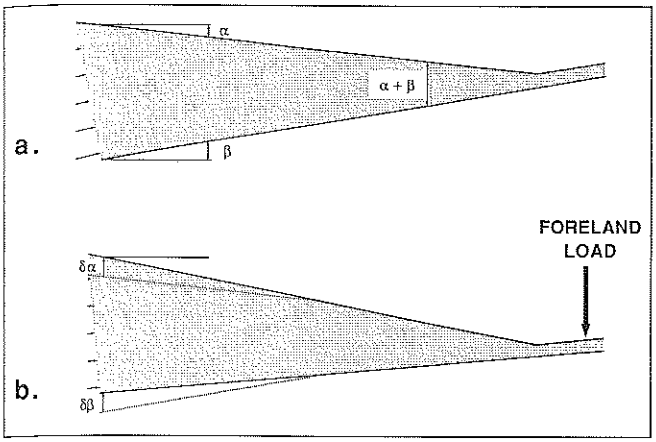


Figure 17. a. Schematic diagram that illustrates the surface slope α and décollement slope β values that define the degree of taper $\alpha + \beta$ of brittlely deforming or stably sliding wedge that is laterally loaded from the thick end and that overlies a weak basal décollement surface. b. Schematic diagram illustrating the change in topographic slope $\delta\alpha$ and décollement dip $\delta\beta$ on the wedge shape of Figure 17a subjected to a foreland flexural load (arrow) with no erosion.

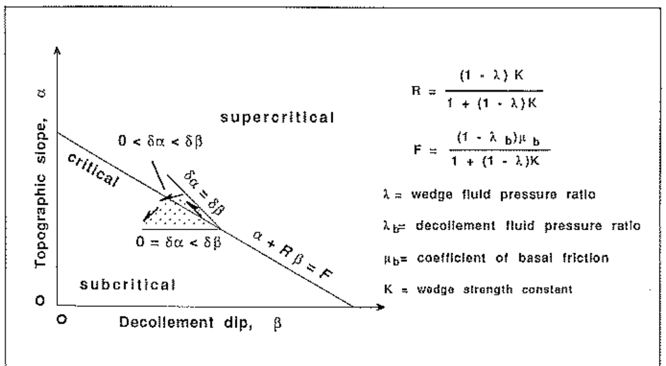


Figure 18. Wedge shape diagram divided by a line of critical wedge shapes into subcritical and supercritical fields of taper. The critical taper equation $\alpha + R\ \beta = F$ relates the topographic slope α and décollement dip β to the fluid pressure properties of the wedge and décollement as well as to the strength of the wedge and the frictional properties of the décollement (see Davis and others, [1983] for derivation). Arrows show the wedge shape path for a model wedge subjected to an instantaneous frontal load and time-dependent erosion. An initially critical wedge can pass through a continuum of supercritical, critical, and subcritical shapes by the combined effects of frontal loading and erosion.

Eocene. The Paleocene subsidence was principally a flexural response to the main phase of uplift for the Wind River and Gros Ventre–Teton basement uplifts to the east and northeast that continued through the early Eocene (Shuster and Steidtmann, 1988). Paleocene and Eocene tectonic subsidence was also demonstrated for the west-central Green River Basin near LaBarge (Shuster and Steidtmann, 1988, Fig. 7), but its relationship to distant foreland uplift loads is more speculative.

The southern margin of the thrust belt experienced similar Paleocene-Eocene subsidence and erosion during uplift of the Uinta Mountains. Figure 19 illustrates the progressive deposition, southeastward tilting, and erosion of the Maastrichtian Hams Fork Conglomerate Member of the Evanston Formation, the late Paleocene upper part of the Evanston Formation, and the early Eocene Wasatch Formation above the thrust belt. The progressive folding and erosion of the upper Evanston and Wasatch formations beneath and adjacent to the Uinta Mountain thrust demonstrate that the erosion and southward tilting of the thrust belt were synchronous with slip on the Uinta Mountain thrust (Fig. 18). The subsidence and deposition, tilting, and erosion of this part of the thrust belt are interpreted to be flexural responses to loading by the Uinta foreland uplift that is analogous to the Paleocene-Eocene flexural subsidence of the northern Green River Basin.

Flexural loads associated with Paleocene to early Eocene emplacement of the Hogsback and Prospect thrust sheets were probably insignificant because erosion of these thrust sheets largely kept pace with uplift (Warner and Royse, 1987). Such erosion is most evident in the LaBarge area where 4.5 km (2.7 mi) of Paleozoic and Mesozoic strata were removed from the Hogsback thrust sheet between late Paleocene emplacement and earliest Eocene overlap of the Wasatch Formation (Fig. 15). The general westward thinning and onlap of Paleocene and early Eocene sediements across the southern thrust belt (Fig. 1) indicate that topographic loads across the belt were insignificant during this period and that the regional subsidence responsible for this onlap was centered to the east in the foreland of the thrust belt.

Flexural subsidence centered in the foreland of a fold-and-thrust belt would reduce the basalt décollement slope (Fig. 17b). In the case in which the reduction in décollement slope $\delta\beta$ is met by an equal increase in topographic slope $\delta\alpha$, the total wedge taper remains unchanged (Fig. 17b). The path $\delta\alpha = \delta\beta$ (Fig. 18) provides one end member of wedge shape that represents instantaneous frontal loading of a wedge without erosion. $0 = \delta\alpha < \delta\beta$ represents the other end member path where uplift of the décollement slope is perfectly matched by erosion and the wedge shape lies entirely in the subcritical field (Fig. 18). Between these two paths lies a field of situations that probably represent the general case for subaerial wedges where $0 < \delta\alpha < \delta\beta$ (shaded on Fig. 18).

If the geometry of a model wedge is tracked through time (arrows on Fig. 18), it is clear that a wedge can pass through a continuum of supercritical through subcritical shapes after foreland subsidence and thrust belt erosion. Because an elastic or viscoelastic lithosphere responds to loading virtually instantaneously (Turcotte and Schubert, 1982), the initial path of a wedge subjected to a frontal load should proceed from an initial critical taper into a supercritical shape approximating the path $\delta\alpha = \delta\beta$. However, erosion rates are time dependent and proportional to relief (Pinet and Souriau, 1988), so the wedge shape would next proceed into the $0 < \delta\alpha < \delta\beta$ field, where it eventually passes through a critical, and eventually subcritical, geometry if the erosion rate approaches the uplift rate. A return to the topographic slope established prior to loading would place the wedge geometry in the subcritical field along the $0 = \delta\alpha < \delta\beta$ path. In addition, flexural or isostatic adjustment from erosion of the topographic load at the rear of the wedge would cause the décollement slope to further decrease, enhancing the subcritical character of the wedge.

The wedge shape path outlined for the frontal flexural load model is offered as a testable hypothesis for frontal slip impedance and internal deformation in the Wyoming-Idaho-Utah thrust belt during the final episode of regional shortening. Unlike most ancient thrust belts, both the décollement slope and the approximate topographic surfaces are well constrained for the late Paleocene and early Eocene of the southern Wyoming-Idaho-Utah thrust belt. In addition, the locations and magnitudes of late Paleocene and early Eocene forelands loads are reasonably contrained, as is the subsidence record of the foreland (Shuster and Steidtmann, 1988).

Foreland stratigraphic and rheologic changes. Mitra and others (1988) noted that the different structural styles of the Wyoming thrust belt and foreland basement uplifts reflect the different physical conditions at the time at which the structures formed. Deformation of the sedimentary cover in the thrust belt generally occurred at greater depths and higher initial temperatures and pressures than in the foreland, largely as the result of the thicker sedimentary section in the thrust belt. The higher temperatures permitted strain-softening mechanisms such as pressure solution slip (Wojtal and Mitra, 1986) to ease thrust slip through the sedimentary cover of the thrust belt. Thicker shale intervals in the Paleozoic rocks of the thrust belt favored the development of regional décollements, particularly where compaction of thick shales helped to maintain high fluid pressures that aid fracture propagation (Atkinson, 1984) and thrust slip (Hubbert and Rubey, 1959). Eastward thinning of important evaporite sequences, such as salt in the Jurassic Preuss Redbeds (Coogan and Yonkee, 1985), could have thwarted thrust slip at higher décollement levels where efficient slip was aided by plastic flow.

The eastward change in stratigraphically controlled rheology is perhaps the most general reason for the cessation of slip along the front of the thrust belt. As physical conditions became less favorable for fracture propagation and basal and internal slip toward the front of the belt, the amount of work required for

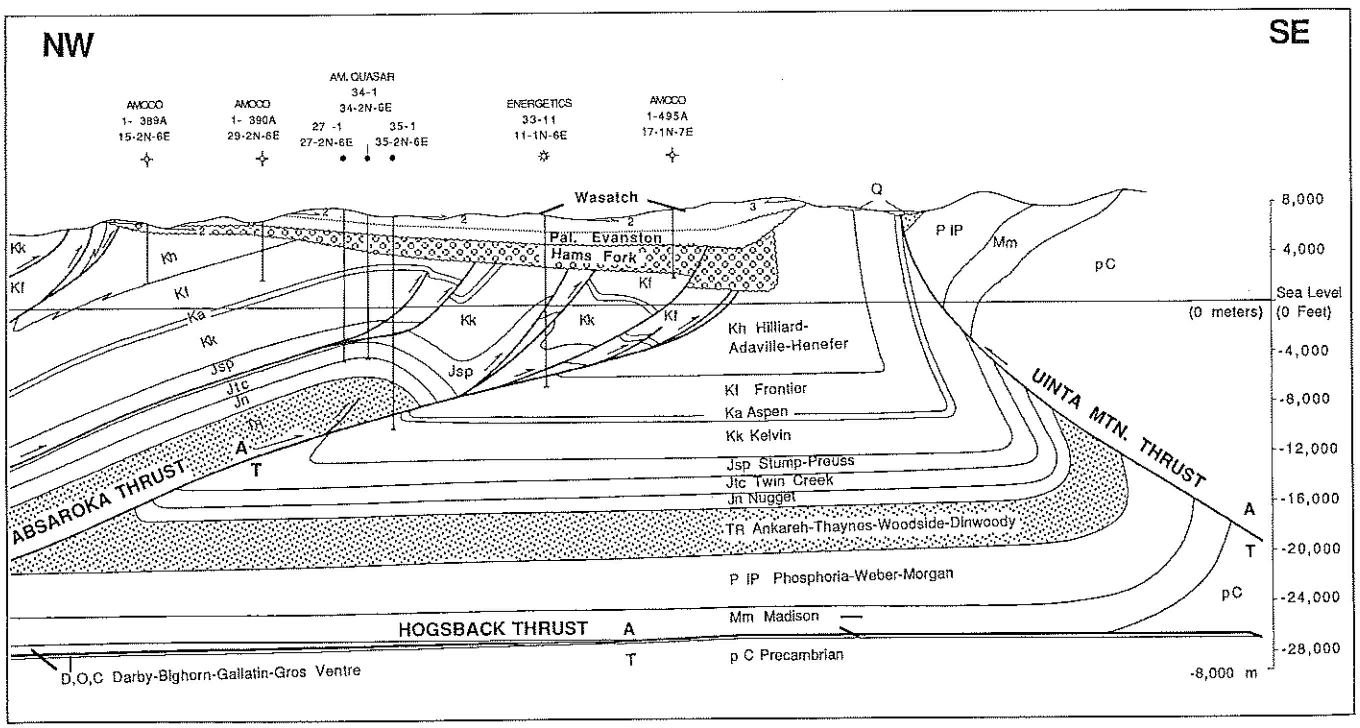


Figure 19. Cross section through southern Fossil Basin, the southern Absaroka thrust, and the north flank of the Uinta Mountains showing the Paleocene to early Eocene subsidence and erosion that accompanied emplacement of the Uinta foreland uplift. The Maastrichtian Hams Fork Conglomerate, the Paleocene Evanston Formation, and the early Eocene Wasatch Formation show southeast dip above the Absaroka thrust sheet. The southeast dip decreases up section and records progressive tilting of the thrust belt during main-phase emplacement of the Uinta uplift. The angular unconformities that bound the Tertiary sequences record continual erosion of the thrust belt during tilting. The cross section is drawn oblique to the transport direction of the Absaroka, Hogsback, and Uinta Mountain thrusts. A = slip component away from viewer; T = slip component toward viewer. Modified from Lamerson (1982, Plate 12 and Figure 22).

further frontal shortening (Mitra and Boyer, 1986) could have exceeded the energy available from the tectonic push applied to the rear of the belt. Break-back reactivation of internal thrusts might have required less work than continued slip along the frontal thrusts or propagation of new thrusts in the foreland. Foreland rheologic changes would also have affected the requirements for stable sliding of the thrust belt according to critical wedge theory, although the specific changes in wedge taper would depend on competing décollement and wedge strength effects. A less efficient basal décollement would require a greater thrust belt taper for stable sliding, whereas a foreland increase in the internal wedge strength would allow stable sliding of a wedge with lower taper (Davis and others, 1983).

CONCLUSIONS

The last phase of shortening in the Wyoming-Idaho-Utah thrust belt was marked by a period of internal thrust reactivation that segmented the top of the regional thrust wedge into a series of piggyback basins between uplifted basin margins. Surface and subsurface structural analysis combined with stratigraphic and sedimentologic study of Bear Lake Plateau basin, northern Fossil Basin, and the LaBarge basin permits the following conclusions:

1. Early Eocene deposition in piggyback basins was controlled by uplift of the basin margins rather than subsidence of basin depocenters.
2. Periods of basin margin uplift are identified by combined observations of (a) alluvial fan sediments shed from the uplifted margins; (b) patterns of onlap, overstep, and angular unconformities along and across the basin margins; and (c) crosscutting structural relations of folded or faulted early Eocene strata along the basin margins.
3. Two sites of basin margin uplift are recognized in the three piggyback basins discussed in this report: Trailing basin margin uplifts occur above the trailing footwall ramps of the thrust sheets that immediately underlie the basins, and leading basin margin uplifts occur above the leading edge of thrust sheets that are underlain by ramps of structurally lower thrusts.
4. Two uplift processes operated in all three basins: reactivated thrust uplift and footwall uplift. Reactivated thrust uplift was generally confined to the trailing ramp areas of thrust sheets as illustrated by uplift of Sublette anticline above the Crawford trailing ramp, and the last phase of uplift above the Hogsback trailing ramp. Footwall uplift was confined to the leading basin margins, where the basin margins and the leading edge of the thrust plate beneath them were translated up the ramp of a structurally lower thrust. Uplift along the leading edges of the Crawford, Absaroka, and Hoigsback thrust sheets was accommodated by uplift along the underlying Absaroka, Hogsback, and LaBarge thrust ramps respectively.
5. Correlation and comparison of times of uplift between the trailing ramp and leading edge areas of a single thrust sheet are important for distinguishing periods when through-going slip is possible. Since thrust slip is communicated from the basal décollement, up the trailing ramp, and finally across the frontal décollement to the thrust front, a cessation of uplift in the trailing ramp region precludes any slip along the frontal length of the thrust. As a corollary, thrust slip along a thrust front requires correlative slip along the trailing length of the thrust.
6. The change from a forward imbrication thrust sequence to internal thrusting during the final episode of thrust shortening was likely the result of impeded slip along the frontal thrusts of the Wyoming-Idaho-Utah thrust belt. Further shortening of the belt was accommodated by reactivated slip along mechanically interconnected thrusts in the interior of the thrust belt. The Crawford, Absaroka, Hogsback-Darby, and Prospect thrusts are mechanically interconnected along the basal décollement in the Cambrian Gros Ventre Formation.
7. Probable causes of slip impedance include foreland buttressing, fault deflection above foreland basement warps, regional wedge adjustment to subcritical taper, and foreland stratigraphic and rheologic changes. The relative contribution of each of the four slip-impeding processes proposed here probably varied in scale and location. On a subregional scale, foeland buttressing was dominant in the northern thrust belt where the frontal thrusts abut the Gros Ventre–Teton and northern Moxa Arch foreland basement uplifts. Similarly, thrust deflection above basement warps was limited to the Moxa Arch near LaBarge in the central thrust belt. On a larger scale, impedance of frontal slip could have been caused by a reduction of the regional wedge taper as a result of the combined effects of subsidence in the foreland and erosion in the thrust belt. However, stratigraphic and associated rheologic changes toward the foreland provide the broadest explanations for impedance of frontal thrust belt slip on a regional scale.

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My interest in the latest Cretaceous and early Eocene stratigraphy and sedimentation of the thrust belt began when I tagged along on a field trip led by Steve Oriel through Fossil Basin and the LaBarge basin in 1985. Prior to that time I considered the "cover" sequence to be a nuisance that only served to obscure the structural geology of the region. I had no idea that I would later retrace Steve's route while trying to recall his words about unconformities and diamictites. We are all fortunate that he rigorously

documented his observations and ideas in print. I began this work after Steve's death, so I never had the opportunity to discuss my interpretations with him. I'm certain that he would strenuously debate some of the ideas presented here. Steve was a demanding taskmaster. He required that one formulate a hypothesis, test it in the field, and document the results in the office. He accomplished all three tasks throughout his career. During our later conversations in Denver, it was clear that Steve regarded the encouragement and guidance of young geologists as an integral part of his work. He spoke with a paternal affection of the many graduate students he had helped along the way. I will always appreciate the kind patience with which he would follow an argument, correct a misconceived notion, and then point to a straighter path toward a solution. This paper is dedicated to Steven S. Oriel.

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