Fold-accommodation faults

Shankar Mitra

ABSTRACT

Fold-accommodation faults are secondary faults that accommodate strain variations related to structural and stratigraphic position during fold evolution. Four main types of fold-accommodation faults are commonly found. Out-of-syncline and into-anticline thrusts form primarily because of an increase in bed curvature within fold cores, although differential layer-parallel strain at different scales also contributes to fault slip. Depending on the kinematic evolution of the major fold, the thrusts may propagate along the steep or gentle limb of an asymmetric fold or along the hinge of symmetric folds. Wedge thrusts are primarily formed in competent units because of variations in penetrative layer-parallel strain between adjacent units. Limb wedges occur as hanging-wall and/or footwall fault-bend and fault-tip folds, whereas hinge wedges occur as multiple nested faults that tend to thicken the more competent units. Forelimb and backlimb thrusts form by a variety of mechanisms. Forelimb space-accommodation thrusts are low-displacement thrusts that resolve strain discontinuities resulting from increased curvature in fold cores. Forelimb shear thrusts form in the late stages of folding because of rotation and layer-parallel extension on the steep forelimbs of folds. Most backlimb thrusts originate as out-of-syncline thrusts. They may eventually link with forelimb thrusts to form forelimb-backlimb thrusts. Back thrusts accommodate hanging-wall strain during the formation of fault-related folds. They either form selectively in competent units or propagate through the section at the same rate as the main thrust. Although fold-accommodation faults are secondary features, they are important elements that define the geometry and size of structural traps in fold-thrust structures. Accurate mapping of these structures is therefore critical in interpreting the structural geometry of fold and thrust belts.

INTRODUCTION

Recent efforts to understand fold-fault relationships in fold and thrust belts have resulted in several geometric and kinematic models of fault-related folds. These models have been based on field observations and kinematic and experimental models to explain the observed features associated with each structural style. As a result,

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models of fault-bend folds (Suppe, 1985; Jamison, 1987; Mitra, 1992), fault-tip or fault-propagation folds (Elliott, 1977; Suppe and Medwedeff, 1984, 1990; Boyer, 1986; Chester and Chester, 1990; Mitra, 1990; Erslev, 1991; Erslev and Mayborn, 1997; Hardy and Ford, 1997; Almendinger, 1998), detachment folds (Jamison, 1987; Dahlstrom, 1990; Mitra, 1992), and duplexes and imbricate thrust systems (Boyer and Elliott, 1982; Suppe, 1985; Mitra, 1986, 1992) are now widely applied to interpret structures with poor or limited data. The common characteristic of all of these structures is that the propagation of thrust faults is instrumental in controlling the geometry and kinematics of the folds.

A second class of fold-fault structures includes those in which the faults are secondary to folds. In this case, the faults accommodate strain variations related to structural and stratigraphic position during folding. These faults are referred to as fold-accommodation faults in this article. Fold-accommodation faults commonly have less slip than major fold-forming faults; furthermore, the slip may vary up and down the dip of the fault, so that the faults commonly terminate within a structure without connecting to a major detachment. Examples of different types of fold-accommodation faults have been described for more than a century (Buxtorf, 1916; Heim, 1919; Willis and Willis, 1934; Douglas, 1958; De Sitter, 1964; Price, 1965; Faill, 1969; Dahlstrom, 1970; Faill and Wells, 1974; Perry, 1975, 1978; Serra, 1977), but their kinematic evolution has not been investigated in detail.

Although fold-accommodation faults are secondary structures, they can involve significant amounts of slip where associated with first-order structures. These faults can, in turn, produce relatively large, lower order folds. Therefore, an understanding of the kinematic evolution and slip distribution of these faults is important for developing accurate interpretations of structures. Models of these structures are applicable to constructing balanced cross sections through fold belts and interpreting the geometry of hydrocarbon trap-forming structures. Furthermore, the slip distribution on some of these faults is important in determining the sealing capacity of fault-related traps.

Most folds are associated with a variety of primary and secondary thrust and normal faults. In this article, we focus on faults that have a similar trend as the major structures and therefore influence the cross-structural evolution of the structures. Transverse faults, which influence the strike-parallel evolution of the structure, are not addressed.

Dahlstrom (1970) documented the terminology for some types of secondary faults on the basis of the structural position of the faults and their vergence with respect to major folds. The more common types of fold-accommodation faults discussed in this article are summarized in Figure 1. Out-of-syncline thrusts originate in the hinges of synclines and propagate through the front limbs or backlimbs of folds. Into-anticline thrusts described in this article form by the same mechanism as out-of-syncline thrusts, except that they terminate in the hinges of anticlines. Wedge thrusts form on the limbs or in the hinges of folds, selectively duplicating some units. Forelimb thrusts cut through the forelimb of an anticline, whereas backlimb thrusts cut through the backlimb. Forelimb-backlimb thrusts alternately transect the forelimbs and backlimbs of folds in a fold train. Cross-crestal faults displace the crest of the anticline in the direction of fold vergence and commonly form by the same mechanism as backlimb or forelimb thrusts. Back thrusts (Serra, 1977) form on the backlimbs of folds and have an opposite vergence to that of the main fold and related thrust.

Most of these secondary faults accommodate (1) changes in the macroscopic structural geometry with structural position and (2) differential penetrative strain between stratigraphic units. This article discusses the evolution of different types of fold-accommodation faults and the influence of these two factors on their kinematic evolution. The slip distribution on the faults and their relationship to major folds is examined in detail. Finally, documented field and subsurface examples of these structures are presented. The conceptual models and examples can be used to interpret the geometry and slip distribution on faults in areas with poor or incomplete data.

To understand the geometric and kinematic relationships between folds and faults, we use models of fold and thrust geometries that use kink-band geometries. These models enable the derivation of key relationships between fold and fault geometries and the analysis of slip distributions between flexural-slip bedding-plane surfaces and faults. It should be recognized that these models represent approximations of folds that may have more concentric geometries. As shown by Woodward et al. (1985), concentric geometries can be approximated using large numbers of kink bands, and curved faults can be approximated using several straight fault segments. The modification and use of these models to simulate complex natural structures is left to the reader.

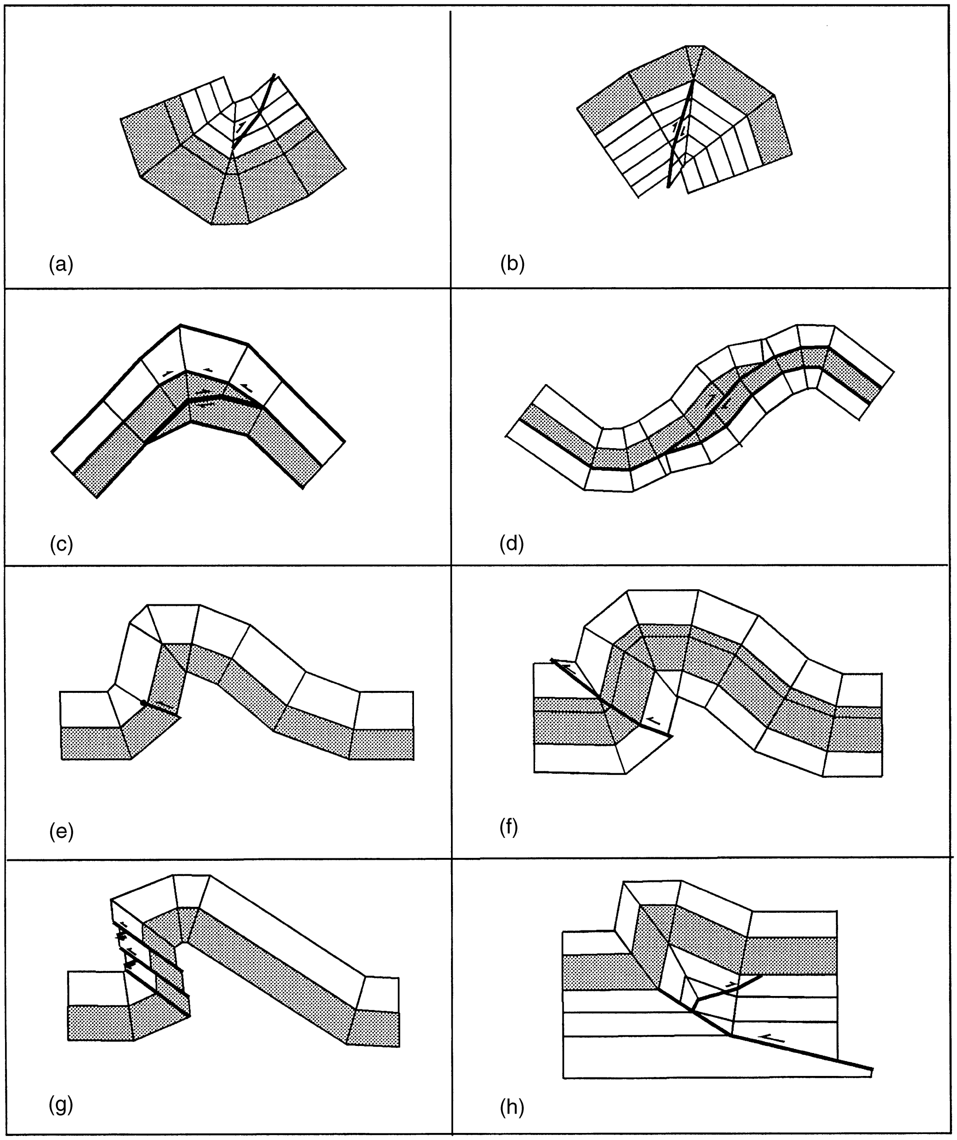


Figure 1. Some common types of fold accommodation faults: (a) out-of-syncline thrust propagating on the gently dipping limb; (b) into-anticline thrust propagating on the gently dipping limb; (c) hinge wedge thrust; (d) limb wedge thrust; (e) forelimb space-accommodation thrust; (f) forelimb-backlimb thrust; (g) forelimb shear thrusts; (h) back thrust.

OUT-OF-SYNCLINE AND INTO-ANTICLINE THRUSTS

Out-of-syncline thrusts originate in synclinal cores and transfer slip along the hinge zone or to the forelimbs or backlimbs of structures. Some of these faults transfer slip to bedding-plane detachments, whereas others lose their slip through penetrative deformation within incompetent units. Price (1965) coined the term "flexural-slip thrust faults," suggesting that these faults are kinematically related to flexural-slip folds and have the same sense of slip as that on bedding planes. In other words, stratigraphically higher units are thrust toward the anticlinal hinge and away from the synclinal hinge. Price (1965) also proposed that these faults form where the curvature of a folded bed approaches a critical limit defined by the thickness of the bed.

Out-of-syncline thrusts may form by three main mechanisms. The most important mechanism is an increase in curvature toward the core of the syncline. Additional components of fault slip may be derived from differential penetrative strain between units and from the formation of secondary structures or disharmonic folds in one or more units.

Thrust faults that decrease in slip and die out near the anticlinal hinge are referred to here as into-anticline thrusts. These thrusts are believed to form by the same mechanism as out-of-syncline thrusts. Again, the thrust may propagate within the hinge area or on the limbs, depending on the mechanism of formation.

First, let us examine the formation of out-of-syncline thrusts due to curvature changes within a fold. Folds are characterized by an increase in curvature or the convergence of axial planes toward the core. This increase in curvature is documented by the convergence and decimation of axial planes in angular folds and by a reduction in the number of arcs representing the fold form in concentric folds.

For an angular syncline (Figure 2a), the total shear (s) for a bed of thickness t, resulting from a change in dip of $\phi_1 + \phi_2$ across the synclinal hinge is given by (Suppe, 1985; Mitra and Namson, 1989)

$$s = 2t[\tan(\phi_1/2) + \tan(\phi_2/2)] \quad (1)$$

This shear is balanced by an opposite sense of shear on the anticlinal arc of a full-wavelength fold, so that the net shear is 0, provided all units in the fold contain the same number of axial surfaces.

At the termination of an axial plane there is a discontinuity in the rate of shear with bed thickness (Figure 2a). The differential shear (Δs) is given by

$$\Delta s = 2t[\tan(\phi_1/2) + \tan(\phi_2/2) - \tan(\phi_{12}/2)] \quad (2)$$

where t represents the thickness of beds from the point of axial convergence; ϕ_1, ϕ_2, and ϕ_{12} represent the change in the dip across respective bends; and $\phi_{12} = \phi_1 + \phi_2$.

The shear results in a reduced length of the beds in the core, unless the shear is exactly balanced by an identical reduction of axial surfaces in other beds on the opposite arc. Except in the case of a perfectly frictionless flexural slip system, such balancing of the axial surfaces is unlikely. Therefore, the termination of axial planes results in a progressive decrease in bed length toward the core of the structure. This is compensated for by the formation of a fault, which originates at the point of discontinuity and increases slip away from the core (Figure 2b).

If the average of the footwall and hanging-wall cutoff angles of this fault is given by β, then the displacement on the fault (d) required to balance the section is given by

$$d = \Delta s \sec \beta \quad (3)$$

In the case of a symmetrical fold, $\phi_1 = \phi_2 = \phi_{12}/2$, and equation 2 reduces to the form

$$\Delta s = 2t[2 \tan(\phi_{12}/4) - \tan(\phi_{12}/2)] \quad (4)$$

Therefore, the normalized differential shear ($\Delta s/t$) is a function of the dip change (ϕ_{12}) or the interlimb angle, given by $\pi - \phi_{12}$. Figure 2d shows a plot of the normalized excess shear vs. the dip change (ϕ_{12}) for a symmetrical fold (curve 1). We see that $\Delta s/t$ is relatively small for open folds but increases dramatically once the interlimb angle falls below 80° ($\phi_{12} > 100°$). Curve 2 shows the case for asymmetric folds for which one of the dip changes (ϕ_1 or ϕ_2) is 40° higher than the other. We see that despite the strong asymmetry, the value of $\Delta s/t$ is not significantly altered. Therefore, the overall dip change or interlimb angle provides a good measure of the differential shear and the fault displacement.

Deviations from the curves may be explained in two main ways. One explanation is that once initiated, these thrusts increase in displacement by replacing bedding-plane slip as the important deformation mechanism. A second factor that may contribute slip to the thrusts is differential penetrative strain between units. The differential strain may be at the microscopic or minor scale, or it may occur by the selective formation of secondary faults or folds within some units. Significantly higher differential penetrative strain between the units may either increase or decrease slip on the thrust. In this case, equation 2 is modified to the form

$$\Delta s = 2t[\tan(\phi_1/2) + \tan(\phi_2/2) - \tan(\phi_{12}/2)] + \Delta l_{ba} \quad (5)$$

where Δl_{ba} is the differential penetrative shortening between units b and a.

A similar analysis may be applied to concentric folds. If the concentric fold form is defined by a series of arcs with radii of curvature $r_1, r_2, r_3, \ldots$ and internal angles (in radians) of $\theta_1, \theta_2, \theta_3, \ldots$, the total shear for a unit of thickness t is given by (Figure 2b)

$$s = t \Sigma \theta_x \quad (6)$$

This shear is exactly balanced by an opposite sense of shear on the anticlinal arc. The termination of an arc results in a cuspate and nonconcentric fold form, with the differential shear of any unit given by

$$\Delta s = t\theta_i \quad (7)$$

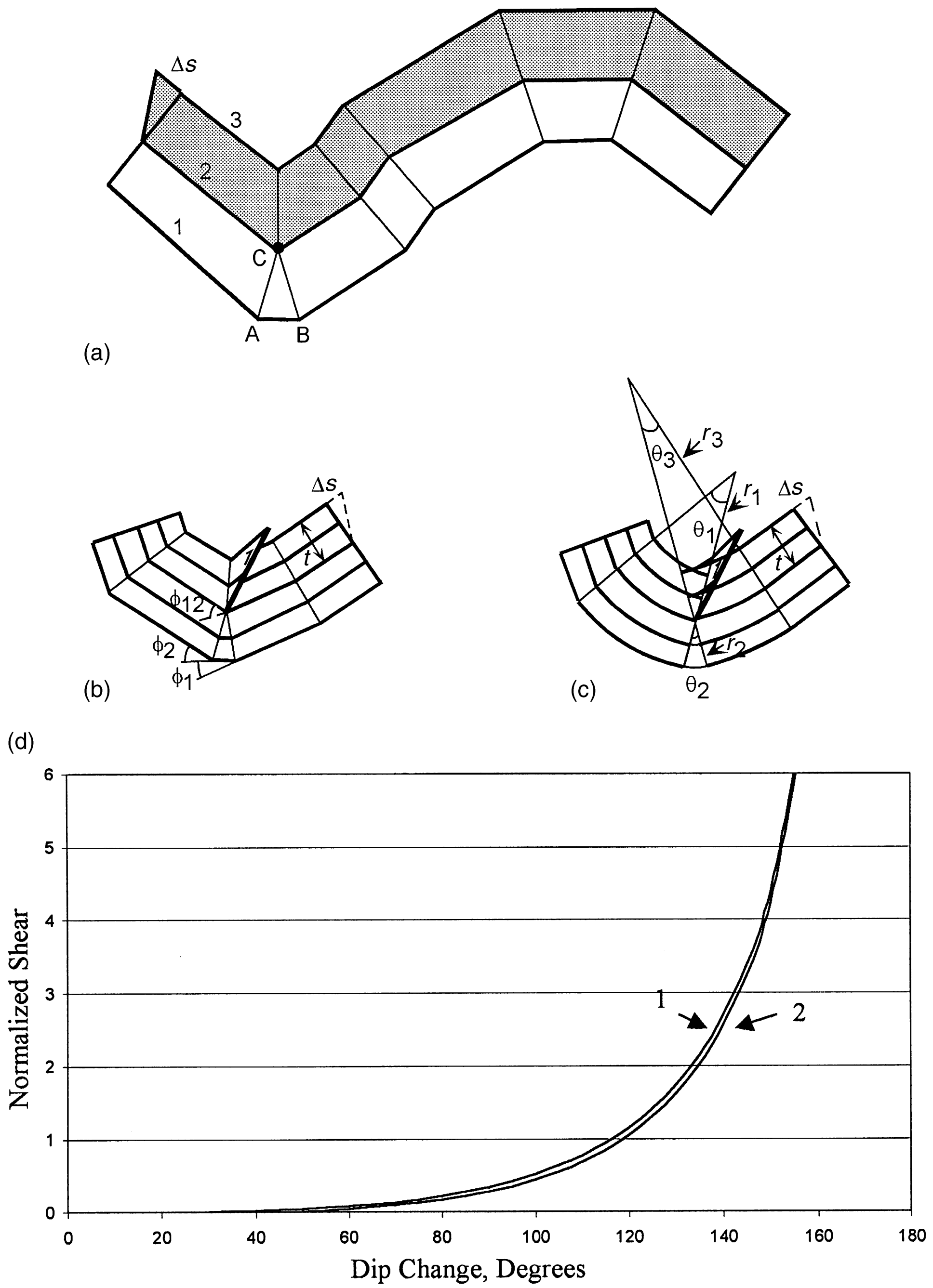


Figure 2. Mechanism of formation of an out-of-syncline thrust due to increased curvature in the fold core. (a) Full-wavelength anticline-syncline pair, with the shear between units 1 and 2 in the syncline balanced by an opposite sense of shear in the anticline. Convergence of axial planes A and B to form axial plane C results in a differential shear between beds 2 and 3. (b) For an angular fold, convergence of axial planes with associated dip changes of ϕ_1 and ϕ_2 results in a differential shear Δs. This shear is resolved by the formation of a thrust, which increases in slip upward and terminates at the intersection of the converging axial planes. (c) For a concentric fold, the termination of an arc with an internal angle of θ_2 results in a shear Δs and a cuspate geometry of the upper units. The shear is accommodated by the formation of an out-of-syncline thrust. The differential shear Δs shown in this and in subsequent figures does not include the homogeneous shear resulting from flexural-slip folding. (d) Graph showing the relationship between the total change in dip and the normalized shear that must be accommodated by out-of-syncline thrusting, assuming that the thrust slip is entirely related to the change in curvature. Normalized shear is the shear differential shear(Δs) divided by the thickness of strata (t) above the point of convergence of two axial planes. Curve 1 represents a symmetrical fold in which $\phi_1 = \phi_2$. Curve 2 represents $\phi_1 = \phi_2 + 40$. The shear is not significantly affected by fold asymmetry.

where θ_i represents the internal angle of the arc that is terminated, and t represents the thickness of the beds between the point of termination and the unit being measured.

This increase in curvature results in a differential bed length, which is compensated for by thrusting of material away from the constricted hinge region (Figure 2c). If multiple arcs are terminated, the total differential shear is given by

$$\Delta s = t \Sigma \theta_i \quad (8)$$

where $\Sigma \theta_i$ represents the curvature of all arcs that terminate between the beds.

Price (1965) proposed that out-of-syncline or flexural slip faults develop in concentric folds where the radius of curvature in the hinge zone is less than the thickness of the bed. This observation agrees very well with the analysis in this article, because the convergence of radii of curvature or axial surfaces would tend to occur where the radii of curvature approach the thickness of a bed.

The exact evolution of out-of-syncline thrusts depends on the kinematic evolution of the fold, determined by the migration of axial surfaces relative to fixed material points. Migration of axial surfaces toward a fixed synclinal hinge may result in faults that are located along the hinge. Migration of axial planes from the steep limb toward the synclinal hinge may result in thrusts that propagate up the steep limb, whereas migration of axial planes from the gently dipping limb toward the synclinal hinge may result in thrusts that propagate up the gentle limb. The location of propagation of the fault may also be governed by other factors, such as preexisting zones of weakness or areas or bedding-plane surfaces characterized by resistance to flexural slip.

We first consider a case where the fold tightens by migration of axial planes from the steep limb of a fold to the hinge zone. The outer arc of the fold in Figure 3a contains four axial planes, whereas all units above the point AB contain three axial planes. Convergence of two of the axial planes, A and B, to one axial plane, AB, results in a higher curvature near the core of the fold and a space problem manifested by a decreased length of units that can be accommodated in the core. The discontinuous shear resulting from the termination of the axial plane is accommodated by the formation of an out-of-syncline fault (Figure 3a).

The slip on the fault at any instant decreases down section, and the fault dies out in the synclinal hinge.

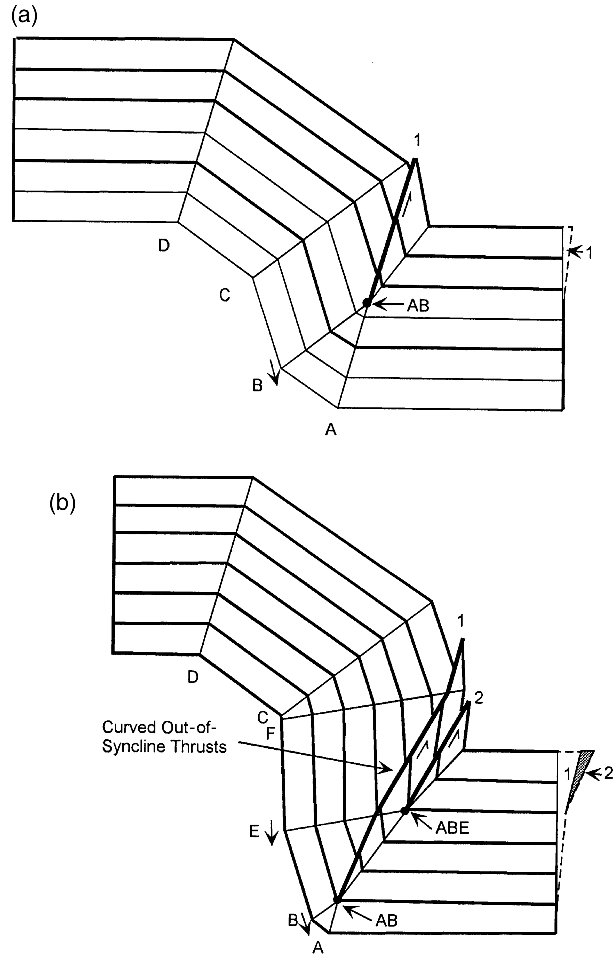


Figure 3. Formation of out-of-syncline thrust on the steep limb of a syncline. (a) Stage 1. Shear related to the convergence of the two axial planes A and B is resolved by the formation of a new axial plane AB and a thrust (1), which terminates at the intersection of A and B. The thrust forms on the steep limb because of the migration of B toward A. The thrust tracks the position of the migrating hinge. (b) Stage 2. Continued fold tightening results in the formation of a pair of new axial planes E and F. Intersection of E with AB results in a new shear and the development of a new axial plane ABE and a second thrust (2). The preexisting thrust continues to propagate into the lower units and is also folded by axial planes E and F. The shear profiles accommodated by the two thrusts are shown. Arrows in this figure and Figure 4 show the direction of migration of the axial planes.

Progressive fold tightening occurs by migration of axial planes on the steep limb toward the synclinal hinge. Assuming that the fault originates at the synclinal hinge, progressive tightening of the fold results in the migration of the synclinal hinge onto the steep limb for the upper units (Figure 3b). The out-of-syncline thrust tracks the position of the fold hinge with progressive deformation, so that the loci of previous hinge points occur progressively updip on the steep limb. The thrust fault is also rotated by newly formed axial planes, resulting in the curved or sigmoidal geometries commonly observed in field examples. New axial intersections (for example, of AB and E to form ABE) result in the formation of multiple subparallel faults (Figure 3b). Although the faults in this example are shown to track the locus of the hinge during progressive deformation, other trajectories of out-of-syncline thrusts are also possible.

Movement of axial planes from the gently dipping limb toward the synclinal hinge results in the migration of the synclinal hinge onto the gently dipping limb for successively shallower horizons. The resulting out-of-syncline thrust is located on the gentle backlimb of the fold (Figure 4a). In the example shown in Figure 4a, the thrust does not track the position of the hinge with time. Backlimb thrust faults with low cutoff angles, such as those documented by Douglas (1958), probably form by this mechanism.

Figure 4b shows the formation of out-of-syncline thrusts associated with migration of two axial planes toward the hinge zone. The locus of the bisecting axial plane does not change direction with progressive deformation (Figure 4b). The out-of-syncline fault forms at a high angle to bedding in the synclinal hinge and retains this position with progressive fold tightening. The slip on the fault increases in the upper units and is accommodated within a narrow zone in the hinge region.

The late Paleozoic units of the Pennsylvania Valley and Ridge are deformed into folds of different orders. Several of these folds are associated with out-of-syncline thrusts. Some examples of these structures are described in the following sections.

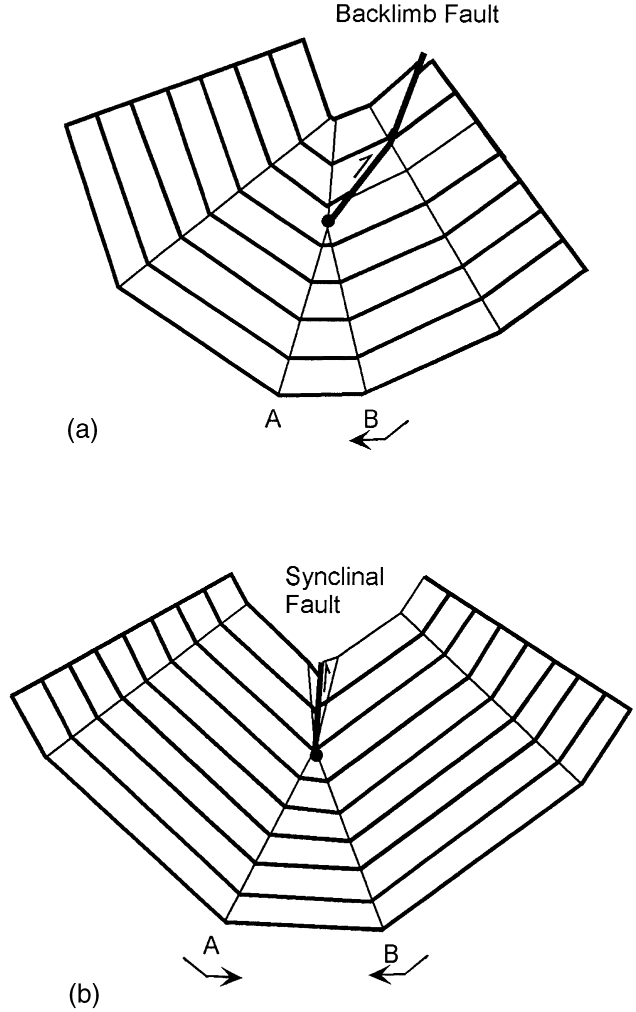


Figure 4. (a) Out-of-syncline thrust located on the gently dipping limb. Axial plane B is migrating toward A. The thrust does not track the position of the migrating hinge in the example shown. (b) Out-of-syncline thrust propagating along the synclinal hinge. Axial planes A and B are migrating toward each other, resulting in symmetric folding and thrust propagation.

Dalmatia Syncline

The Tuscarora anticlinorium is one of the major first-order folds in the Valley and Ridge. Some third-order folds related to a second-order fold at the hinge of the Tuscarora anticlinorium are exposed in the Devonian Mahantango Formation. One such third-order syncline in the Dalmatia Member is exposed in a quarry on the south limb of the second-order fold (Figure 5). The bedding planes, shown in dashed lines in Figure 5, indicate upward-tightening of the hinge. Two major out-of-syncline thrusts that terminate near the synclinal hinge can be mapped. A large number of secondary thrusts propagating from bedding planes are also present.

Mount Pleasant Mills Syncline

Folded Devonian–Pennsylvanian units in the Valley and Ridge fold belt in Pennsylvania show several good examples of out of the syncline flexural-slip thrusts. Figure 6a shows a map (modified from Faill and Wells, 1974) through a syncline near Mount Pleasant Mills, in the Millerstown quadrangle. The folded units include the Devonian Marcellus and Mahantango formations. Faill and Wells (1974) mapped a fault on the gentle northern limb of the anticline that originates near the top of the Devonian Marcellus shales and increases slip up section as it cuts through units into the lower units of the Devonian Mahantango Formation.

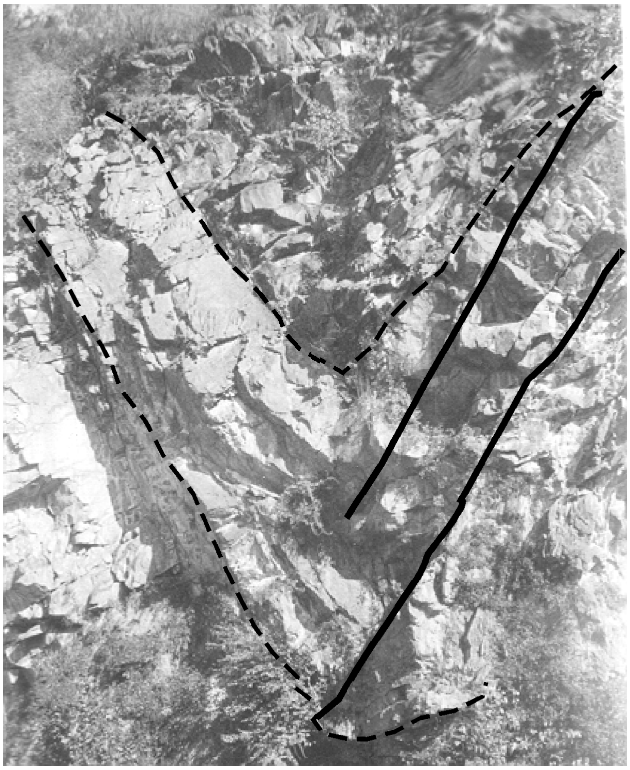


Figure 5. Out-of-syncline thrusts within sandstones of the Dalmatia Member of the Devonian Mahantango Formation in one of the Dalmatia quarries in the Pennsylvania Valley and Ridge. The third-order syncline is located on the south limb of a second-order anticline near the hinge of the Tuscarora anticline. The bedding planes (dashed lines) indicate upward tightening of the hinge. Numerous additional small-scale thrusts are also present.

An interpretive cross section through this structure is shown in Figure 6b and illustrates the increase in slip up section. The out-of-syncline thrust is related to the tightening due to increased curvature in the core of the syncline. Additional components of slip may be related to penetrative deformation within the Marcellus shales and the formation of secondary folds on the limbs of the structure (Figure 6b). One such secondary fold is located on the north limb of the structure and is restricted to the Turkey Ridge member of the Devonian Mahantango Formation and lower stratigraphic units. A second structure occurs on the steep north limb of the syncline within the Devonian Marcellus Shale and lower units.

Kuh-e Dashtak–Kuh-e Shah Nishin Syncline, Zagros Fold Belt

The Zagros Mountains folded belt has been characterized as a classic example of a concentrically folded belt. The Cretaceous and younger units are characterized by broad anticlines and pinched and narrow synclines. Out-of-syncline thrusts should be common within these pinched synclines, particularly in the more competent units. Figure 7 shows an example of an out-of-syncline thrust between Kuh-e Dashtak and Kuh-e Shah Nishin described by Colman-Sadd (1978). This thrust, which is exposed at the surface, terminates within limestones of the Lower–Upper Cretaceous Bangestan Group. The throw on the fault increases through the Upper Cretaceous and Eocene marls of the Gurpi and Pabdeh formations and the Oligocene–Miocene limestones of the Asmari Formation before decreasing within the incompetent Miocene Gachsaran Formation, where flowage by penetrative deformation accommodates some of the fault slip.

The thrust propagates along the synclinal hinge, suggesting possible symmetrical axial plane migration toward the hinge of the fold during the formation of the thrust fault. The additional folding related to the propagation of the out-of-syncline thrust is confined to a narrow zone in the hinge of the structure.

WEDGE THRUSTS

Flexural-slip folding occurs by slip on bedding planes, with the slip on the outer arc directed toward the anticlinal fold hinge. Bedding-plane thrusts periodically climb section through competent units, transferring slip to other bedding-plane thrusts within a relatively incompetent unit (Cloos, 1964). The sense of shear on the wedge thrusts is commonly congruous with that associated with flexural-slip folding. Wedge thrusts with an incongruent sense of slip compared to the fold are commonly not synchronous with the folding. Although many wedge thrusts form on fold limbs (limb wedges), folds also contain wedge thrusts located in the hinge zone (hinge wedges).

Hinge Wedges

The relative abundance of wedge thrusts in fold hinges is probably related to the fact that incompetent units typically exhibit significant thickening and layer-parallel shortening in the hinge zones. This shortening must be accommodated by faults in the more competent units.

Consider a subangular fold containing a shale and a sandstone unit (Figure 8). The formation of a parallel fold results in a homogeneous shear. Layer-parallel

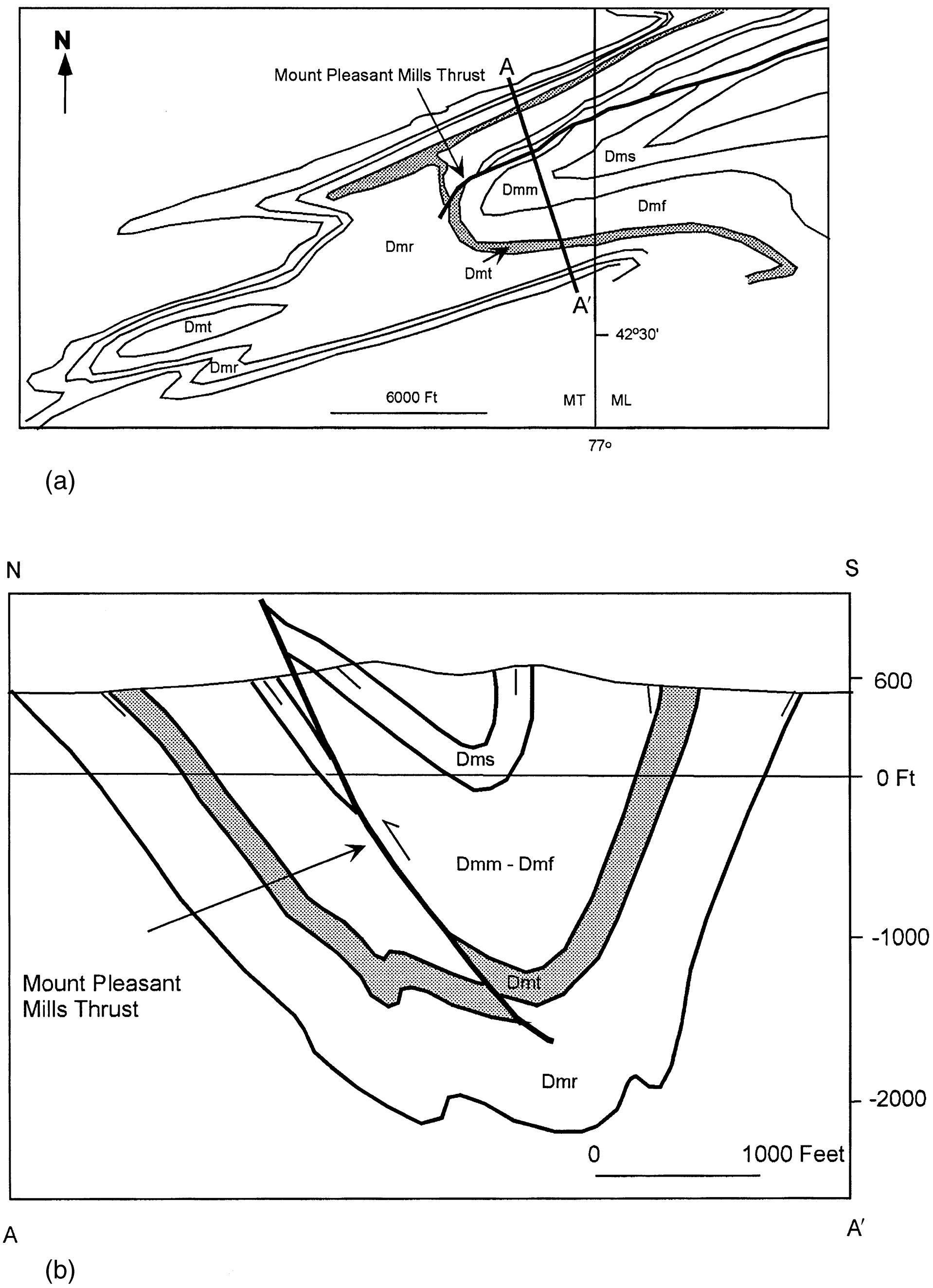


Figure 6. (a) Geological map of the Mount Pleasant Mills syncline in the Millerstown and Millersburg quadrangles in the Pennsylvania Valley and Ridge (modified from Faill and Wells [1974] and Hoskins [1976]). The Mount Pleasant Mills thrust is located on the gently dipping north limb of the syncline. Line AA′ shows the approximate location of the cross section in part b. Dmr = Middle Devonian Marcellus Formation; Dm = Devonian Mahantango Formation, which includes the following members: Dmt = Turkey Ridge; Dmf = Fisher Ridge; Dmm = Montebello; Dms = Sherman Ridge.
(b) Structural cross section through the Mount Pleasant Mills syncline. The Mount Pleasant Mills thrust is interpreted as an out-of-syncline thrust formed primarily because of increased curvature in the core of the syncline. The slip on the thrust decreases down section, and the thrust terminates in the Dmr units. Some component of slip on the thrust is also related to the occurrence of secondary folds, primarily in the Dmr and Dmt units.

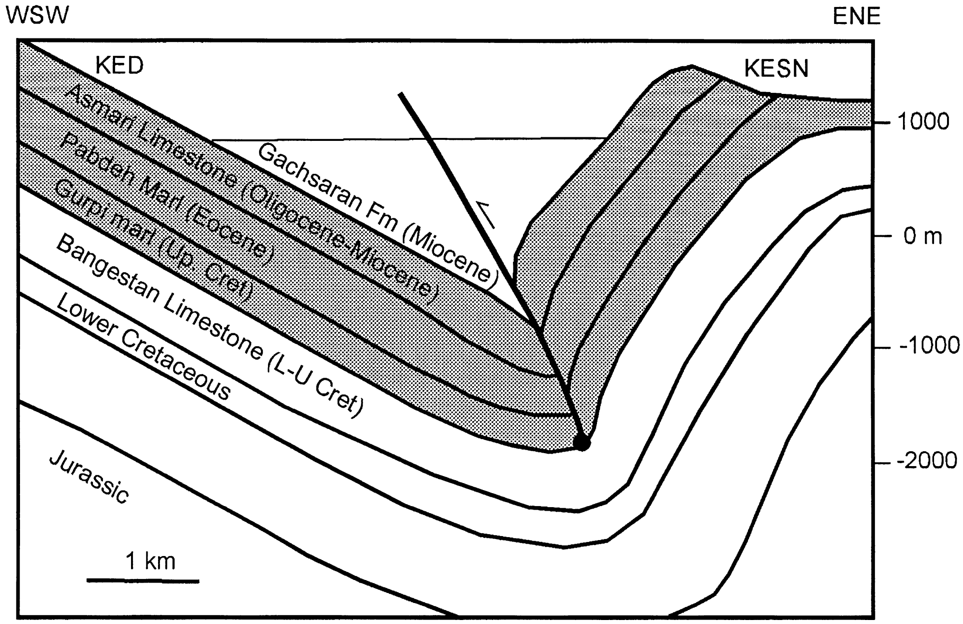
shortening within the shale unit is confined to the hinge region and results in an increase in thickness to a new thickness t'. The excess layer-parallel shortening within the shale unit (Δl_{sh}) is given by

$$\Delta l_{sh} = A/t \qquad (9)$$

where A is the excess area within the shale unit. The sandstone unit, however, exhibits parallel folding with a constant thickness orthogonal to bedding. This results in constriction of the sandstone unit in the hinge area. Therefore, the additional shortening within the shale unit must be accommodated by a wedge thrust in the sandstone unit, so that $\Delta l_{sh} = \Delta l_{ss}$. The thrust originates as a bedding-plane thrust at the base of the sand unit and transfers slip to another bedding-plane thrust at the top of the unit. Increasing curvature within the fold core may also result in an additional component of fault slip.

Note that wedge faults also form during layer-parallel shortening prior to any significant folding of the beds. The wedge zones subsequently provide

Figure 7. Out-of-syncline thrust propagating along the synclinal axis in the Kuh-e Dashtak (KED)–Kuh-e Shah Nishin (KESN) structure, Zagros fold belt (modified from Colman-Sadd, 1978). The thrust terminates within the Miocene evaporites of the Gachsaran Formation.



nucleation sites for folds (P. Geiser, 2001, personal communication). These early wedge faults can be distinguished from the hinge wedges discussed here by the fact that the thickening of the adjacent incompetent units is likely to be distributed over a larger area.

Pennsylvania Valley and Ridge

Figure 9a shows an example of wedge thrusting in an anticlinal core in interlayered sandstones and shales of the Devonian Catskill Formation from the Pennsylvania Valley and Ridge. The fold has a subangular geometry with thickening of beds in the curved hinge zones. The wedge faults occur as curved surfaces within the red siltstones, whereas the adjacent shales exhibit well-developed cleavage, suggestive of layer-parallel shortening.

A smaller scale example of multiple core wedges is shown in Figure 9b. This structure is in the Silurian Tuscarora Formation, located on the steep northwestern limb of the Jacks Mountain anticlinorium in the Pennsylvania Valley and Ridge. The thin-bedded siltstone at the bottom is folded into a tight fold. The overlying competent sandstones deform by flexural slip, but because of a low density of bedding-plane slip surfaces, the hinge zone contains a large number of

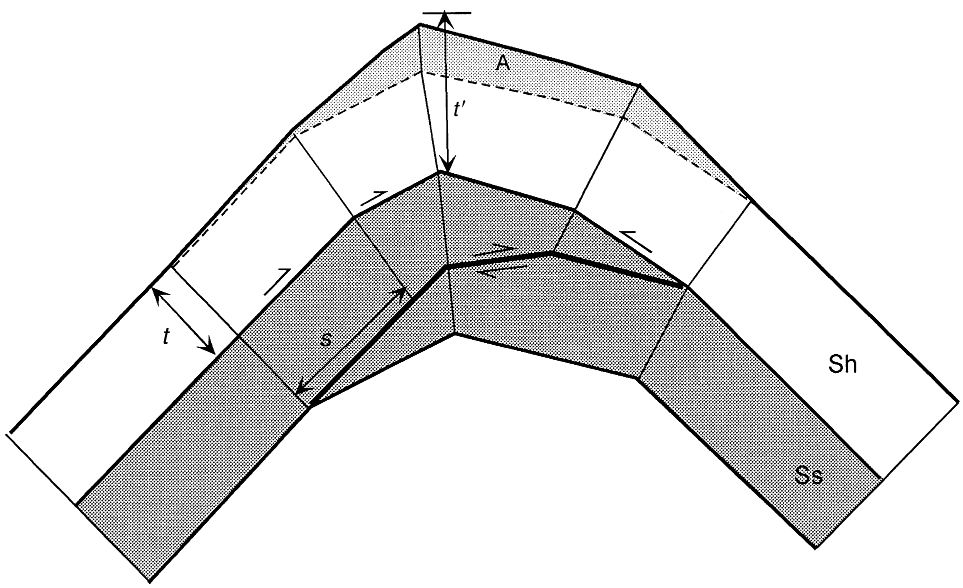
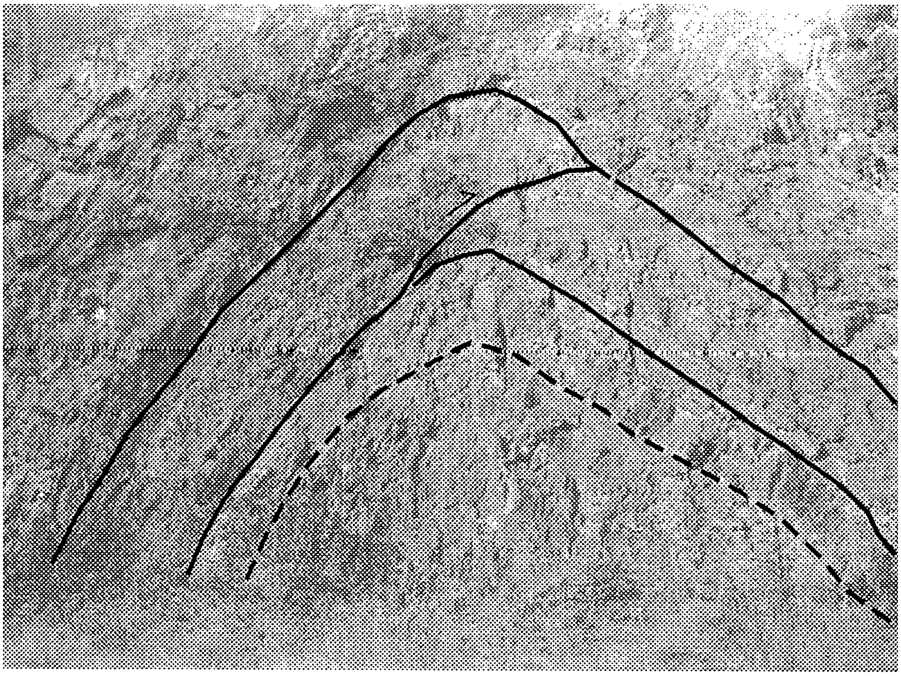
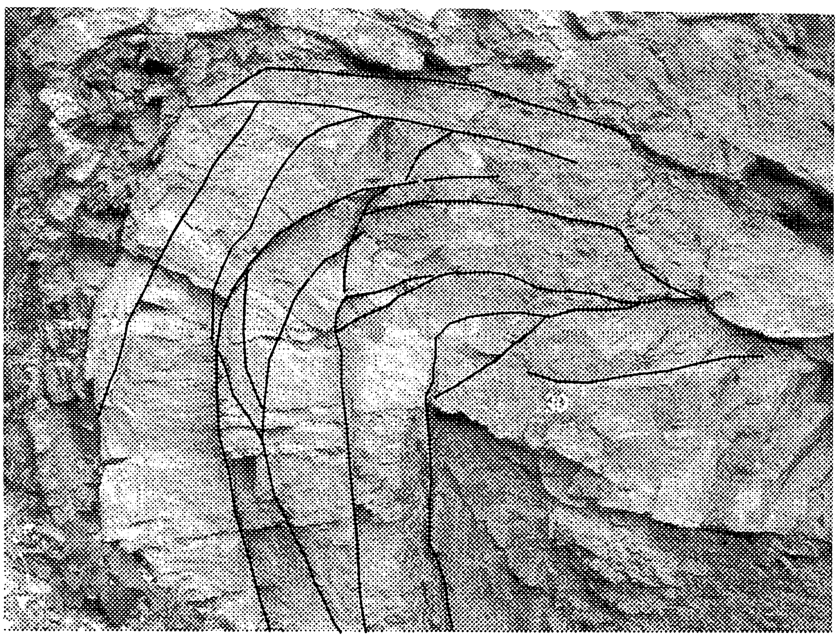


Figure 8. Formation of a wedge thrust in the hinge of an anticline. Shortening associated with slip on the thrust in the competent unit (Ss) is compensated for by thickening of the incompetent unit (Sh) from an original thickness of *t* to a final thickness of *t′* in the hinge region.



(a)

Figure 9. (a) Wedge thrust within the Devonian Catskill Formation from the Pennsylvania Valley and Ridge. The wedge is restricted to the siltstone unit, whereas adjacent shale units are relatively unfaulted. (b) Multiple wedge thrusts in the hinge of a minor fold within the Silurian Tuscarora Formation near the Laurel Creek reservoir in the Pennsylvania Valley and Ridge. Because of the lack of abundant bedding-plane slip surfaces in the sandstone units, folding occurs primarily by the formation of numerous wedge thrusts terminating in discontinuous bedding-plane surfaces.



(b)

minor faults and cataclastic bands. These minor faults transfer slip between the bedding planes, forming overlapping wedges. The net effect of the wedging is significant thickening of the hinge zone in comparison to the limbs.

Ventura Avenue Anticline

The Ventura Avenue anticline in the Ventura basin (Figure 10) is a symmetrical detachment fold located above a basal detachment within Miocene shales (Yeats, 1983; Huftile and Yeats, 1995; Davis et al.,

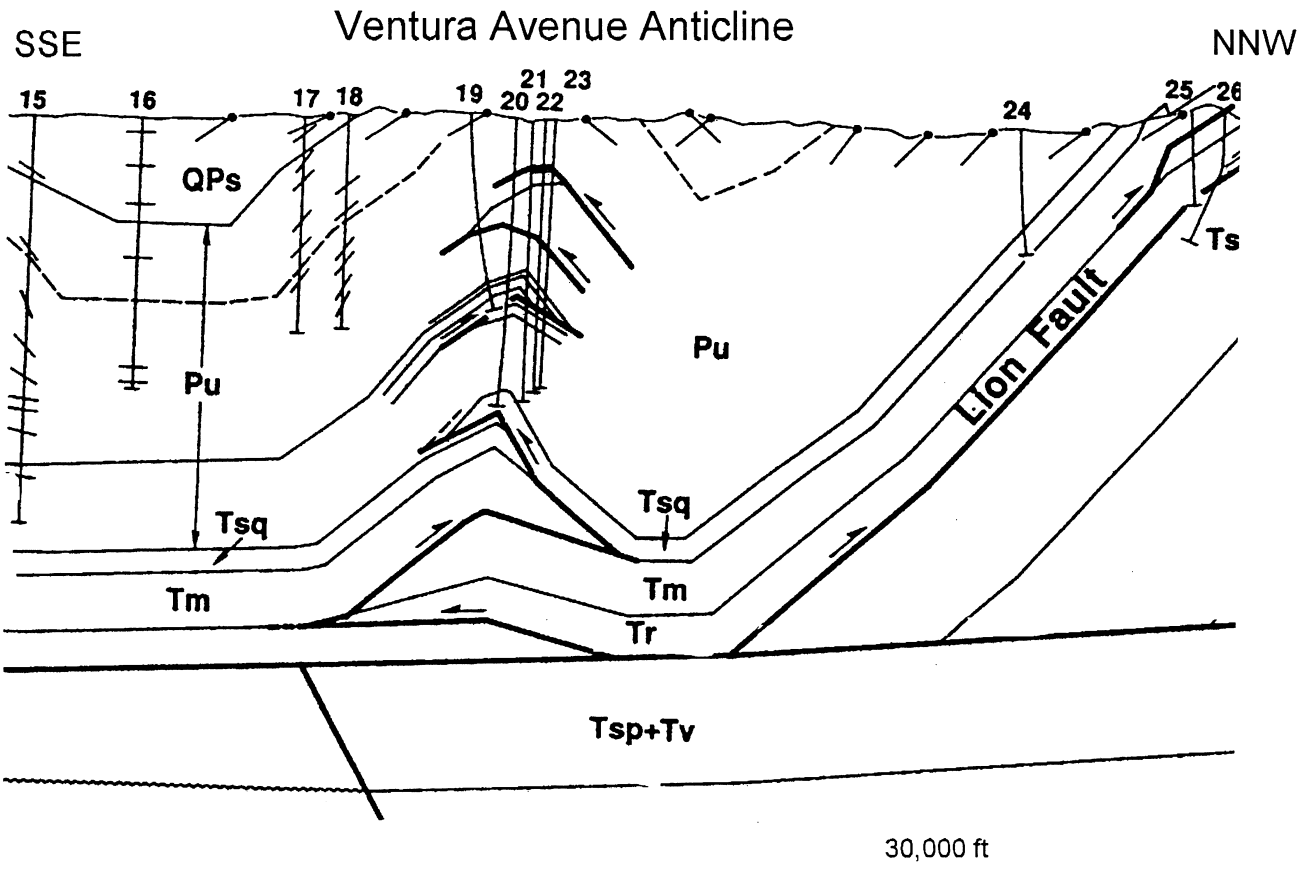


Figure 10. Structural cross section through the Ventura Avenue anticline in the Ventura basin, California (modified from a larger section in Davis et al., 1996). Tsp + Tv = Sespe and Vaqueros formations; Tr = Rincon formation; Tm = Modelo and Monterey formations; Tsq = Sisquoc Formation; Pu = Pico Formation; QPs = undifferentiated Saugus Formation and various alluvial units. Interpretation of well data suggests significant thickening of the Tm, Tsq, and Pu units by wedge thrusting. A similar interpretation of the structure was proposed by Yeats (1983).

1996). The Miocene–Pliocene clastic units, including the Modelo, Monterey, Sisquoc, and Pico formations, show significant thickening in the hinge zone (hinge to limb thickness ratio of approximately 1.8:1). Furthermore, well data from the crestal area suggest repeated duplication of these units along opposite-dipping thrust faults. Davis et al. (1996) have interpreted the hinge zone to consist of a series of at least five major wedge faults, some with displacements exceeding 1 km, whereas Yeats (1983) has interpreted an even greater number of wedges. This structure is an example of a large-scale analog of a fold containing hinge wedges.

Limb Wedges

The primary mechanism for the formation of limb wedges is differential layer-parallel strain between adjacent units, with the wedges forming in the more competent units. Figure 11a shows an example of how the differential strain between two units results in a wedge thrust on a fold limb. Consider an anticlinal-synclinal pair within a multilayer consisting of lower and upper incompetent units (A and C, respectively), and a middle competent unit B. Assume that all three units undergo a component of layer-parallel shortening during folding with $s_a = s_c > s_b$. Then, the shortening associated with thrusting is given by

$$s_t = s_a - s_b \tag{10}$$

Assuming equal area and a uniform distribution of strain over an original length ℓ from the synclinal to the anticlinal hinge, each of the units undergoes layer thickening (Δt) to a new thickness (t'), so that $t'_a/t_a = t'_c/t_c > t'_b/t_b$.

Then

$$t'_a(\ell - s_a) = t_a\ell \tag{11}$$

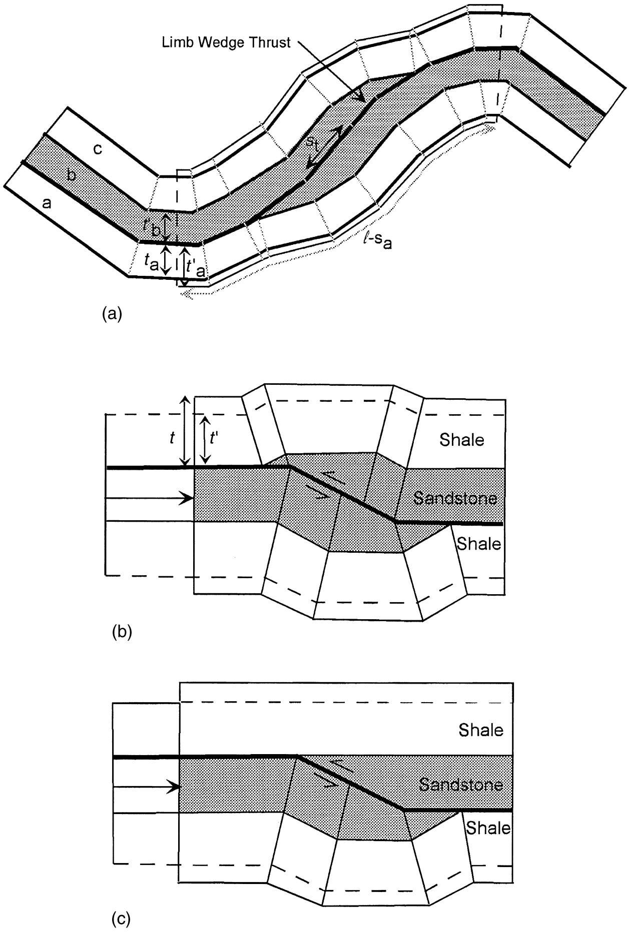


Figure 11. (a) Geometry of a limb-wedge thrust. Wedging of the competent unit b is compensated for by layer-normal thickening of unit a. The wedging results in the formation of a pair of broad secondary anticlines and synclines. (b) Symmetrical limb wedge consisting of anticlinal and synclinal fault-bend folds in a sandstone unit between two incompetent shale units. (c) Asymmetrical limb wedge with a synclinal fault bend in a sandstone unit between two shale units, which deform differentially by layer-parallel shortening.

so that

$$s_a = \ell(1 - t_a/t'_a) \tag{12}$$

and

$$s_b = \ell(1 - t_b/t'_b) \tag{13}$$

Therefore,

$$s_t = s_a - s_b = \ell(t_b/t'_b - t_a/t'_a) \tag{14}$$

If $t'_b = t_b$,

$$s_t = \ell(1 - t_a/t'_a) = \ell(1 - 1/(1 + e_a)) \tag{15}$$

or

$$e_a = [\ell/(\ell - s_t)] - 1 \tag{16}$$

where e_a is the layer normal strain in unit A.

Figure 11a shows the special case where $t_b = t_a$. Similar equations apply, relating the shortening between units B and C. The value of ℓ can be determined by restoring b, and the displacement on the fault (s_t) can be measured directly. Therefore, the layer normal strain in unit a (e_a) can be determined using equation 16.

Limb-wedge thrusts may involve fault-bend folding of both of the fault blocks (Figure 11b), commonly the hanging wall, or one of the blocks, or fault-tip folding with significant loss of slip along the fault ramp (Figure 11c). Significant slip may be transferred by the formation of systems of secondary duplexes. Cloos (1964) documented several examples of wedge thrusts of all of these varieties from the limbs of folds in the central Appalachians.

The Silurian–Pennsylvanian clastic sequence in the Pennsylvania Valley and Ridge is folded into several open anticlines and synclines. Some excellent examples of limb-wedge thrusts have been documented from this area by Faill and Wells (1974). The wedging is commonly confined to the Montebello and Sherman Ridge members of the Devonian Mahantango Formation. Folding of the hanging wall with a stationary footwall is shown in a series of cross sections through the Oriental fault. Gentle folding of both blocks is exhibited by the Devonian Old Port Formation in the Flint Ridge structure.

Buffalo Mountain Fault

Perhaps the best example of a large-scale limb-wedge thrust is the Buffalo Mountain fault found on the gentle northern limb of the Buffalo-Berry synclinorium in the Pennsylvania Valley and Ridge. The structure was mapped by Faill and Wells (1974) and Hoskins (1976). In map view (Figure 12a), the thrust fault originates as a bedding-plane thrust in the Duncannon Member of the Devonian Catskill Formation, which consists of an upward-fining sequence of sandstones, siltstones, and claystones. The thrust cuts up section through the ridge-forming competent Mount Carbon, Beckville, and Spechty Kopf sandstone members of the Devonian–Mississippian Pocono Group and flattens out into siltstones and shales of the Mississippian Mauch Chunk Formation. Note that all units return to their original stratigraphic thickness away from the

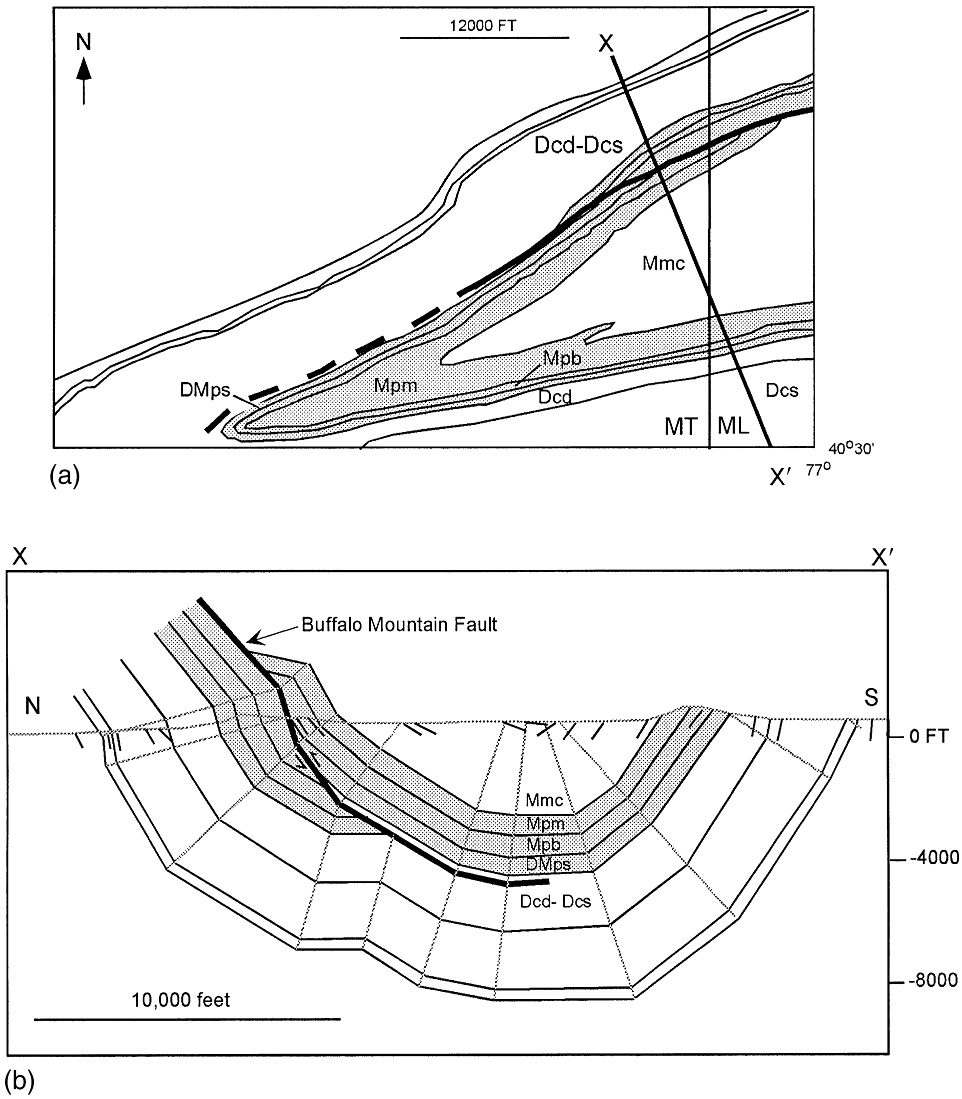


Figure 12. (a) Map of the Buffalo-Berry synclinorium in the Millerstown (MT) and Millersburg (ML) quadrangles in the Pennsylvania Valley and Ridge (modified from Faill and Wells [1974] and Hoskins [1976]). The Buffalo Mountain fault is interpreted to originate at the synclinal hinge and form a wedge fault within the Devonian–Mississippian Pocono Formation. Dcd-Dcs = Duncannon and Sherman Creek members of the Devonian Catskill Formation; DMps = Spechty Kopf Member of the Mississippian Pocono Formation; Mpb and Mpm = Beckville and Mount Carbon members of the Pocono Formation; Mmc = Mauch Chunk Formation. Location of cross section XX′ in part b is shown. (b) Structural cross section XX′ through the Buffalo-Berry synclinorium showing a limb wedge associated with the Buffalo Mountain fault. The fault duplicates sandstones units in the Pocono Formation by forming both anticlinal and synclinal fault-bend folds. The formation of the wedge explains the formation of the Centerville kink bands described by Faill and Wells (1974).

wedge zone. Figure 12b shows a cross section through the structure. The thicknesses of beds and the distribution of dips through the structure clearly indicate a wedge with a folded footwall and hanging wall. Faill and Wells (1974), although recognizing the hanging-wall fold, did not describe wedge-related footwall folding; however, they did recognize a series of kink bands north of the fault within the Catskill Formation, which they referred to as the Centerville kink bands. Faill and Wells (1974)concluded that the distance of these kink bands and their sense of rotation did not suggest a correlation to the Buffalo Mountain fault. Careful examination of these bends, however, indicates that they represent the dip panels associated with footwall folding along the wedge fault. The displacement on the thrust within the wedge is approximately 2200 ft (670 m). Assuming that the Pocono units undergo little penetrative deformation and that deformation within the Catskill and Mauch Chunk formations occurs primarily by layer normal thickening, distributed over the entire length of the arc between the synclinal and anticlinal hinges, the thickening of units to balance the structure is calculated to be less than 20%.

Wedge thrusts may originate as out-of-syncline or into-anticline thrusts, especially if they terminate in the hinges of major synclines or anticlines. The distinctive feature of wedge thrusts is that they result in secondary folds restricted to a relatively narrow zone on the fold limb. Outside this zone, the units return to their original thickness. Furthermore, the geometries and/or structural positions of the secondary folds are closely tied to fault ramps.

FORELIMB AND BACKLIMB THRUSTS

Forelimb and backlimb thrusts are related to folds in two main ways. First, a thrust may migrate from the forelimb to the backlimb of a structure and vice versa (forelimb-backlimb thrust) as it cuts through different stratigraphic units. In this case, either the thrust fault is kinematically related to a larger structure and therefore cuts across the secondary folds, or it formed later than the folds. Second, the thrust fault is confined to a forelimb or backlimb location of a structure, thereby suggesting a kinematic relation with the fold. Subsequent linking of independently developed forelimb and backlimb thrusts can also result in forelimb-backlimb thrusts. The second type of thrust is the more common and is discussed in more detail in a following section.

Several hypotheses (Figure 13) for the formation of forelimb thrusts have been proposed (De Sitter, 1964). Buxtorf (1916) proposed that thrust faults formed early and were subsequently folded passively during the fold formation. Buxtorf's (1916) model was based on studies in the Jura Mountains, where folding above a weak detachment is the primary mechanism of deformation. Heim's (1919) model relates the formation of forelimb thrusts to major nappes in the Swiss Alps. In this model, the front limb is folded to an overturned geometry. Subsequent shearing of this limb results in the development of a ductile deformation zone and the propagation of a thrust fault through the highly attenuated limb. This type of thrust fault is referred to as a stretch thrust because the formation of the thrust fault accommodates stretching of the front limb. This model is most applicable to areas involving low-grade metamorphic rocks, in which significant shear strains may develop through ductile deformation. De Sitter (1964) proposed a model in which an early formed fold is faulted on the front limb, possibly in response to tightening of the structure. The thrust is subsequently folded by ongoing folding. Willis and Willis (1934) proposed the term "break thrust" for faults that break through the deformed front limbs of anticlines. Their model is similar to that of Heim (1919), except that it applies to structures formed under low temperatures and pressures.

Two main reasons exist for the formation of forelimb thrusts, space accommodation and forelimb shear.

Forelimb Space-Accommodation Thrusts

A change in bed curvature across different stratigraphic units results in space problems within the structure. Because the front limb is commonly steep, the area between the anticlinal and synclinal hinges is commonly characterized by the greatest change in curvature. If the curvature increases down section because of the convergence of two axial planes, A and B, to a single axial plane, AB (Figure 14a), space problems in the core of the anticline are resolved by the development of a low-angle thrust on

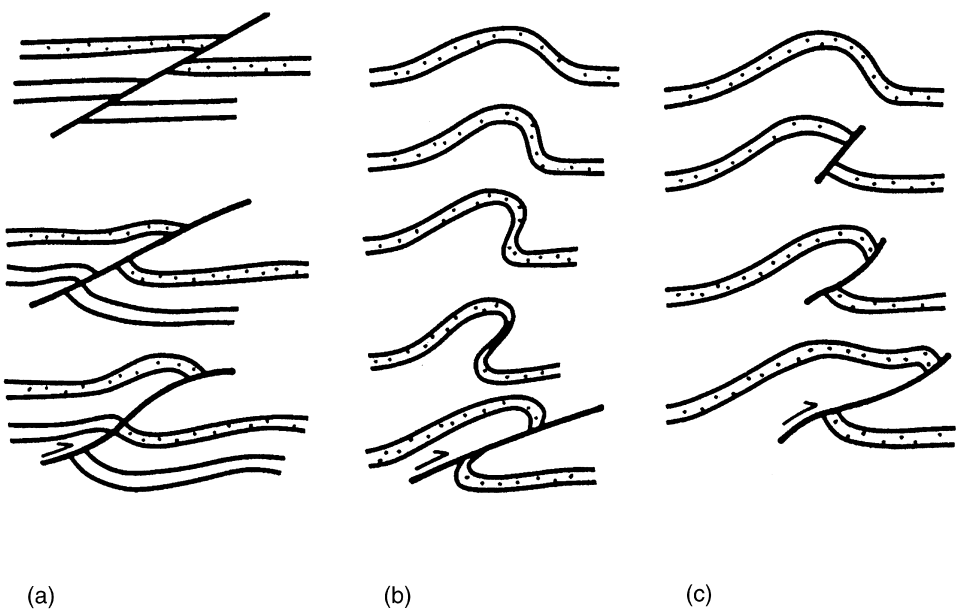


Figure 13. Proposed mechanisms for the formation of forelimb thrusts. (a) Buxtorf's (1916) model suggests the warping of original planar thrusts during later folding. (b) Stretch thrust model proposed by Heim (1919) invokes the propagation of a thrust through the sheared forelimb. (c) DeSitter's (1964) model suggests the formation of a thrust fault on the fold forelimb due to space accommodation. The thrust fault is warped because of continued folding.

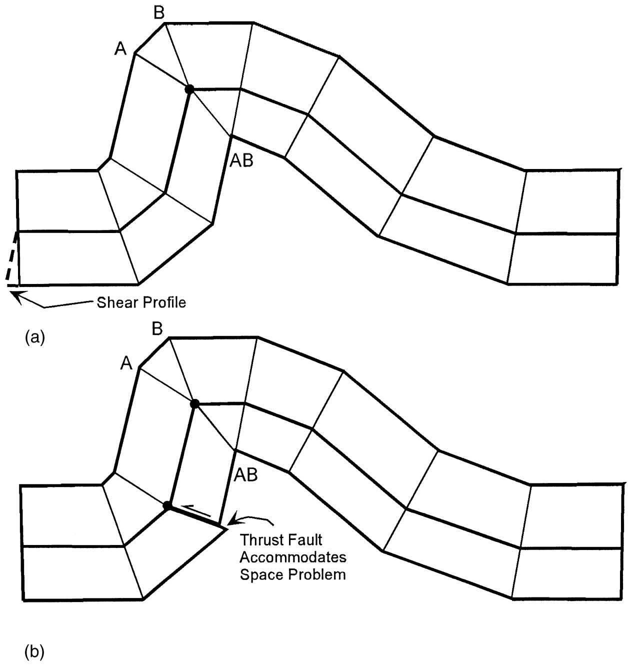


Figure 14. Model for the formation of a forelimb thrust due to space accommodation. The intersection of two axial planes A and B into a single axial plane AB results in a shear within the lower units (a). This shear is accommodated by the formation of a thrust that propagates through the front limb and terminates at the same stratigraphic horizon as the point of intersection of A and B (b).

the front limb (Figure 14b). The mechanism is very similar to that for out-of-syncline and into-anticline thrusts, except that the space problems are accommodated on the front limb rather than in the hinge. The thrust fault loses slip up section and dies out at the stratigraphic horizon marking the convergence of the two axial planes. With increasing tightening of the fold, the point of convergence of the axial planes and the thrust fault propagate up section. This model is similar to that proposed by De Sitter (1964) for the formation of thrusts in the cores of concentric folds.

Forelimb Shear Thrusts

The front limbs of asymmetric folds commonly undergo steepening and rotation in the late stages of folding, possibly due to frictional drag on the thrust ramp or upper detachment. The steepening occurs between locked synclinal and anticlinal hinges. This induces a forward shear on the front limb and, if the front limb is close to vertical, a component of layer-parallel extension (Figure 15). The shear may be accommodated by penetrative deformation, by a system of forelimb thrusts parallel to the direction of shear, or by the formation of a single major fault. Depending on the location of the fault, it may be described as a cross-crestal fault or a forelimb shear fault. The shear faults are commonly associated with secondary conjugate faults. The thrusts may connect down section with a major thrust ramp or die out within incompetent units in the anticlinal core.

Wills Mountain Anticlinorium, Central Appalachians

The front limb of the Wills Mountain anticline in the Pennsylvania, Maryland, and West Virginia Valley and Ridge, contains numerous examples (Figure 16a) of

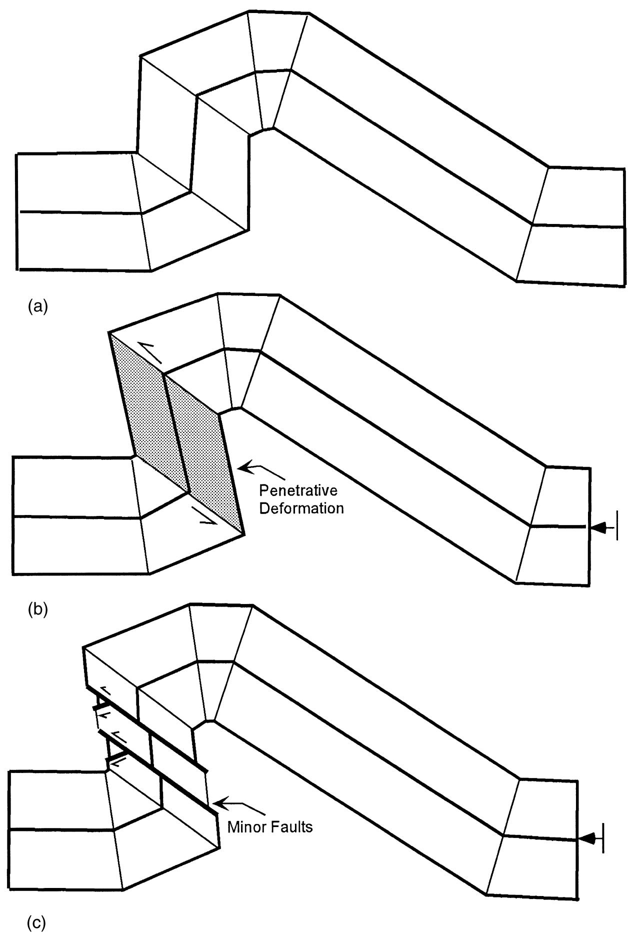


Figure 15. Model for the formation of forelimb shear thrusts. (a) An asymmetric anticline is first formed. (b, c) Continued tightening of the fold with a fixed synclinal axis results in shear on the front limb. This shear may be resolved by penetrative ductile or brittle deformation (b), or the formation of discrete minor faults (c). Secondary faults that are conjugate to these main faults may also form.

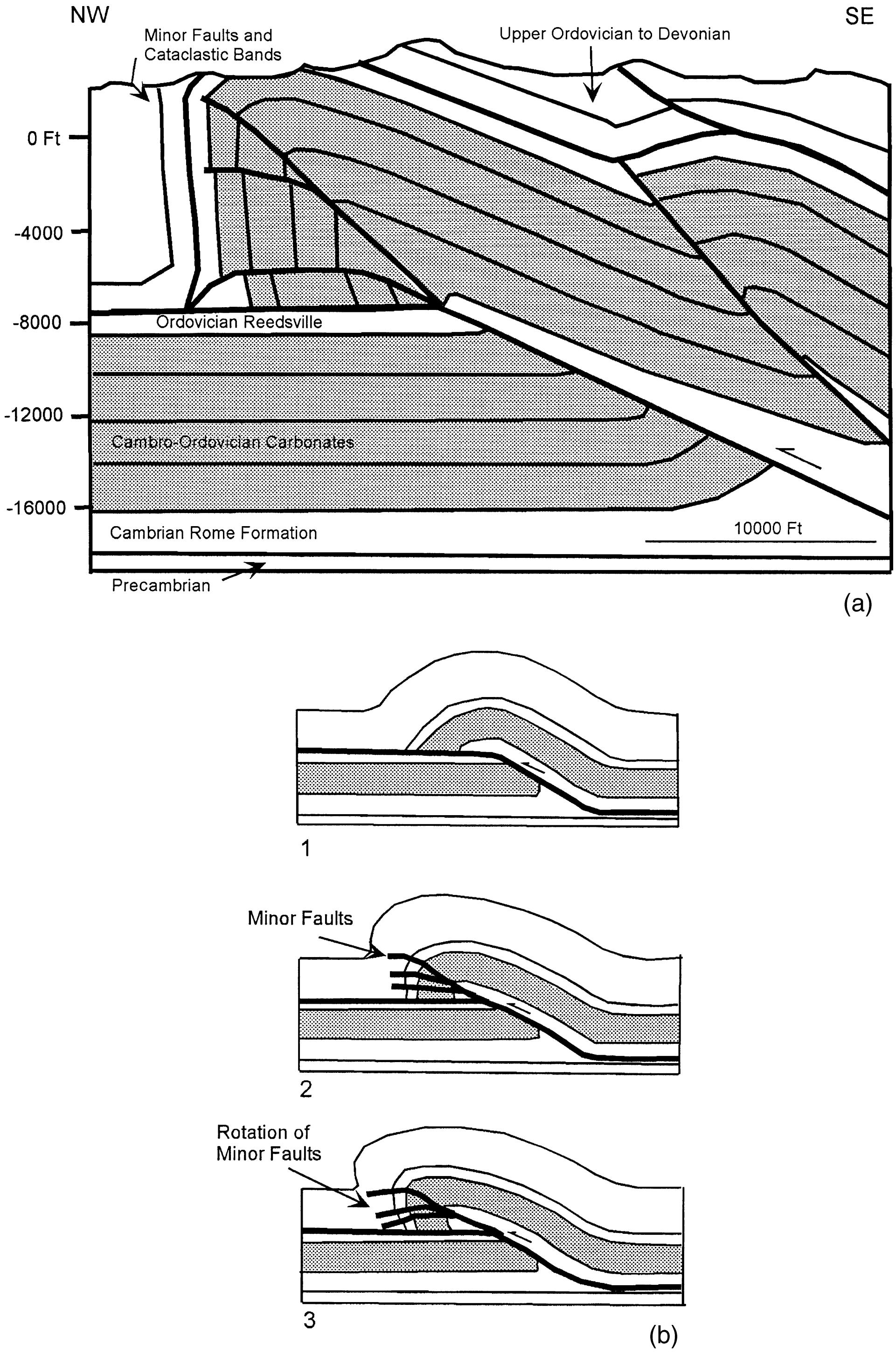


Figure 16. (a) Cross section through the Wills Mountain anticline, Pendleton County, West Virginia Valley and Ridge (modified from Perry, 1978). Note the presence of forelimb shear thrusts on the front limb. (b) Model for the formation of sheared and rotated forelimbs for fault-related structures such as the Wills Mountain anticline.

these structures at different scales (Mitra, 1987). The faults are commonly west dipping because of additional forward rotation of the limb after the formation of the fault (Figure 16b). Perry (1975, 1978) documented examples of conjugate faults in the West Virginia Valley and Ridge, and similar structures were documented by Faill et al. (1973) on the steep limb of the Jacks Mountain anticlinorium.

Puri Anticline, Papua New Guinea

The leading edge of the Papua New Guinea thrust belt is defined by a series of tight fault-tip folds. The Puri anticline is one of the frontal structures in the belt, located close to the southeastern boundary of Papua New Guinea. The structure has been interpreted using seismic reflection data, dipmeter data from the Puri 1 well and two sidetracks (Puri 1A and 1B), and surface data (Medd, 1996). A modified version of the interpretation from Medd (1996) is shown in Figure 17. The larger structure has been interpreted by Medd (1996) to be a triangle zone in which the upper part of the Pliocene Orubadi and Pliocene–Pleistocene Era formations are deformed into a series of north-verging thrusts, whereas the Cretaceous Ieru Formation, the Eocene limestones, the Oligocene–Miocene Puri Formation, and the lower part of the Orubadi Formation are deformed into a tight south-vergent fault-tip fold overlying a series of imbricate thrusts (Medd, 1996, figure 8).

We focus on the structural geometry of the south-vergent structure in the core of the fold. The units are deformed into a tight fault-tip fold. The front limb of the structure is cut by a series of low-angle imbricates, which steepen with depth and also extend bedding. Several of these faults are intersected by the Puri 1 and sidetrack wells. These faults are interpreted to have formed because of forelimb shear in the late stages of folding.

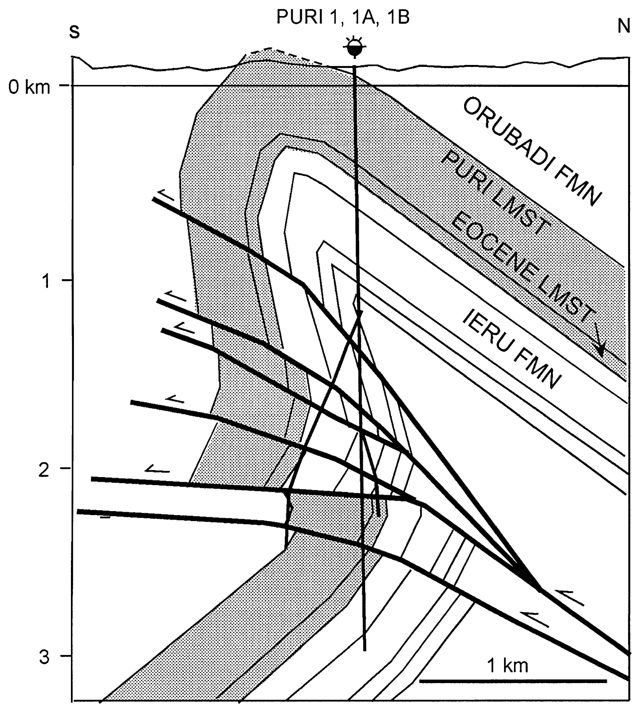


Figure 17. Structural cross section through the Puri anticline, Papua New Guinea fold belt. The cross section is modified based on data in Medd (1996). The front limb of the tight anticline is cut by a series of low-displacement thrust faults that also extend bedding. These thrust faults are interpreted to have developed in the late stages of folding to accommodate forward shear of the front limb.

Forelimb-Backlimb Thrusts

Backlimb thrusts may form as a result of space accommodation and commonly originate as out-of-syncline thrusts. In this case, the slip on the thrust increases upward along the backlimb. With increasing fold tightening, the point of convergence and the thrust fault both migrate down section. Downward-propagating backlimb thrusts may eventually link with upward-migrating forelimb thrusts to form forelimb-backlimb thrusts (Figure 18a). In this case, the slip on the fault is a minimum at the junction and increases away from it. An example of the incipient development of such a thrust is seen in a fold within the Oriskany sandstones in the Smoke Hole area of the West Virginia Valley and Ridge (Figure 18b). Note that the fold contains several thrusts that migrate from the forelimb to the backlimb and vice versa. The forelimb thrust at the bottom left of Figure 18b loses slip up section, whereas a backlimb thrust originates within the syncline. Splay thrusts from each of the two main thrusts link up to form a single forelimb-backlimb thrust.

Backlimb thrusts also form as a result of slip transfer from the main fold-related fault. The back thrust commonly converges with the main thrust at depth, as observed in the Turner Valley anticline in the Alberta Foothills thrust belt (Gallup, 1951). Finally, backlimb thrusts can originate as wedge thrusts to accommodate differential penetrative strain between two beds.

BACK THRUSTS

Antithetic accommodation thrusts, also known as back thrusts, are secondary imbricate thrusts that have an opposite dip and vergence and a conjugate sense of slip to the main thrust. They accommodate deformation of units in the hanging wall. These thrusts typically originate at fault bends or regions of high curvature on faults and accommodate the deformation in the hanging wall associated with movement through fault bends. They may also originate at points with high frictional resistance on the main thrust. Serra (1977) documented several examples of back thrusts and also showed that their origin is in agreement with experimental models and numerical simulations (Barber and Sowers, 1974) of ramp-related folds. Serra's (1977) observations indicated that for ramp-related folds, back thrusts commonly originate at the base of the footwall ramp and propagate into the hanging wall with a convex-upward trajectory.

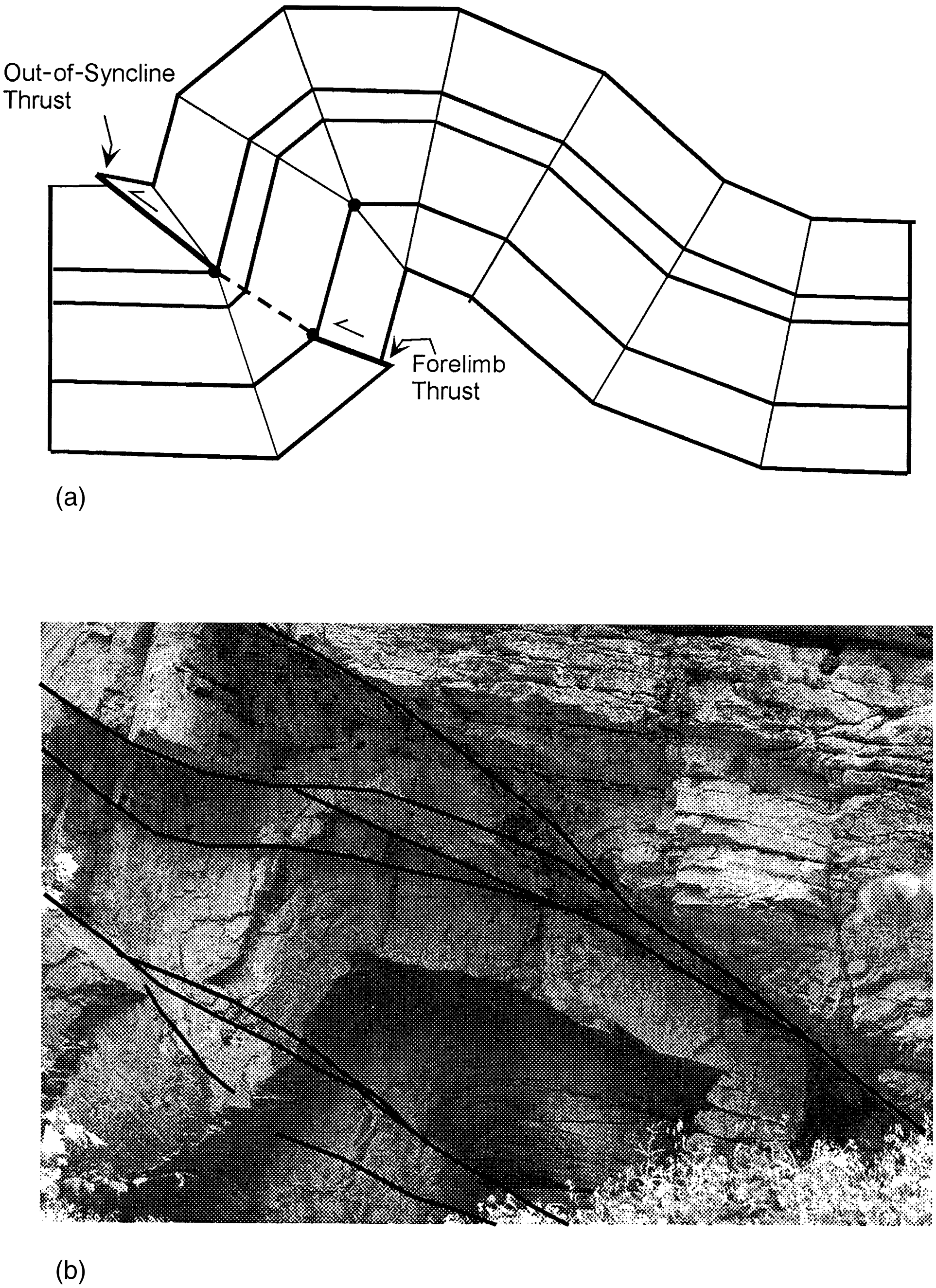


Figure 18. (a) Model for the formation of forelimb-backlimb thrusts. An upward-propagating forelimb thrust and a downward-propagating out-of-syncline thrust form as space-accommodation thrusts. These thrusts may eventually connect along the dashed line to form a forelimb-backlimb thrust. (b) Forelimb-backlimb thrusts within a third-order fold in the Devonian Oriskany Formation in the Smoke Hole area, West Virginia Appalachians. The thrusts at the bottom left suggest splaying and linkage of forelimb and synclinal thrusts.

The most general case of back thrusting involves the accommodation of deformation by folding, faulting, or a combination of the two mechanisms. Deformation of the hanging wall by folding results in the formation of a new dip panel, whose dip depends on the mechanism of deformation in the hanging wall (Figure 19a). At some stage, frictional resistance results in accommodation of deformation by the formation of a back thrust that propagates through the backlimb (Figure 19b). Continued displacement results in the formation of a new back thrust (Figure 19c). Alternatively, the deformation may be accommodated by folding and faulting on the backlimb, with the back thrusts passively transported up the fault ramp. The final point of intersection of each back thrust with the main thrust is always further up the main fault than its point of origin.

The nature and geometry of back thrusts can vary widely, depending on their kinematic origin and the mechanical stratigraphy involved. Two commonly

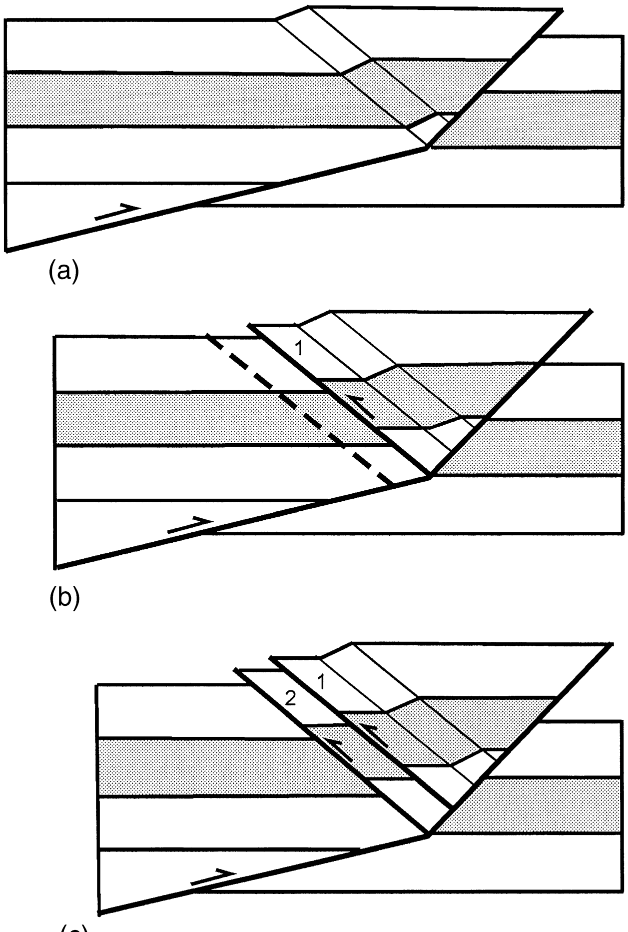


Figure 19. Model for the partitioning of deformation between folding and faulting on the backlimb. (a) Deformation of the hanging wall by folding initially results in the formation of a new dip panel. (b) Frictional resistance associated with fault slip results in accommodation of deformation by the formation of a back thrust that propagates through the backlimb. (c) Increasing deformation results in the formation of a new back thrust.

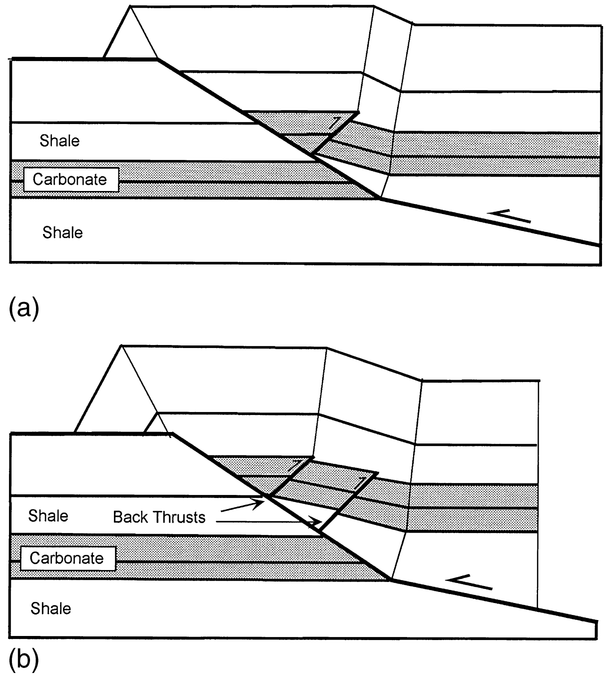


Figure 20. Evolution of back thrusts within a competent unit in a fault-bend fold. The formation of the back thrust results from only partial rotation of the carbonate unit as it passes through the synclinal fault bend. These back thrusts are either confined to or dominantly developed in the competent units. They have a periodic spacing and accommodate the additional strain associated with rotation through the fault bend. Some of the examples of back thrusts documented by Serra (1977) are of this form.

observed types of back thrusts are discussed here: (1) those confined to a competent stratigraphic unit between two relatively incompetent units deforming primarily by fault-bend folding, and (2) those formed during intense folding within thin-bedded units of relatively uniform lithology, commonly associated with fault-propagation or fault-tip folding.

The first type of back thrust occurs within a competent unit being folded through a fault bend (Figure 20). The competent unit is only partially rotated as it passes through the synclinal fault bend, resulting in the additional strain being accommodated by the formation of a back thrust. The back thrust is either confined to or dominantly developed within the competent unit. Continued movement on the main fault results in the formation of multiple faults with a periodic spacing. Some of the examples of back thrusts documented by Serra (1977) are of this form.

Back thrusts associated with fault-propagation folds in thin-bedded strata are typically folded with progressive deformation. Consider a fold within a propagating thrust above a fault bend. Movement of units through the bend in the fault results in the formation of a back thrust. The back thrust also separates a zone of tight folding above from a relatively unfolded zone below. In the particular case shown in Figure 21, the backlimb dip also increases from below the back thrust to above it, so that the synclinal axial plane on the backlimb steepens in units not cut by the back thrust. Both the main thrust and the back thrust propagate upward with increasing slip (Figure 21b). If the front limb steepens with increasing fault slip as proposed by Mitra (1990), the back thrust is folded on the

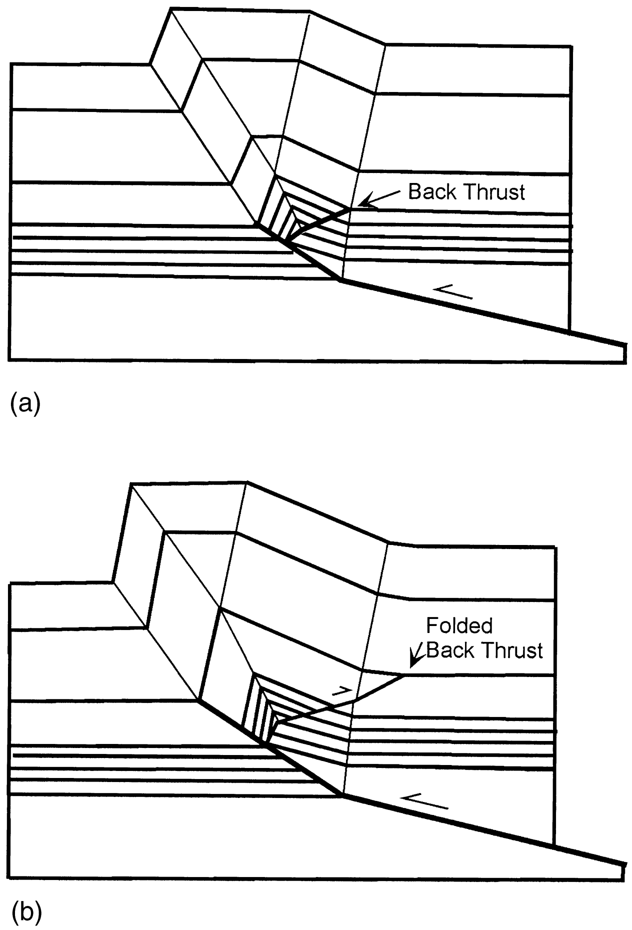


Figure 21. Evolution of a back thrust in homogeneous thinly bedded units within a fault-propagation fold. The back thrust originates at the synclinal bend and typically propagates up section at a similar rate to that of the major fault (a). Note the refraction of the synclinal axial plane between the faulted and unfaulted units (b). The thrust is also rotated to a steeper dip on the front limb of the fold because of limb rotation and on the backlimb of the fold because of folding by the synclinal axis.

steep front limb, accounting for the convex-upward geometry observed for many back thrusts. Furthermore, with increasing fault propagation, the back thrust eventually cuts the synclinal axial surface on the backlimb of the fold and is subsequently folded into a concave-upward geometry. Therefore, the final geometry of the back thrust is sigmoidal, with steeper dips on the forelimb and in the unfolded region behind the synclinal axis (Figure 21b).

Dunlap Fold, Tennessee Plateau

Figure 22 shows a third-order fault-propagation fold through the Dunlap section in the Appalachian Plateau in Tennessee. The main thrust terminates close to the top left of Figure 22. The backlimb of the structure contains several back thrusts. The most prominent of these originates on the main thrust at the boundary between the thin-bedded shales and a sandstone unit. The back thrust is rotated to a steeper dip on the front limb of the fold and by the synclinal axis on the backlimb.

CONCLUSIONS

Fold-accommodation faults are secondary faults that accommodate strain discontinuities during folding. The discontinuities may be related to variations in structural or stratigraphic position. Studies of maps, cross sections, and field examples of these structures at different scales have been used to determine the kinematic evolution of these faults. Some key characteristics of these faults are that they commonly have smaller slip than fold-forming faults and that the slip can vary significantly updip or downdip. As a result, these faults commonly terminate within a structure without being connected to a major detachment.

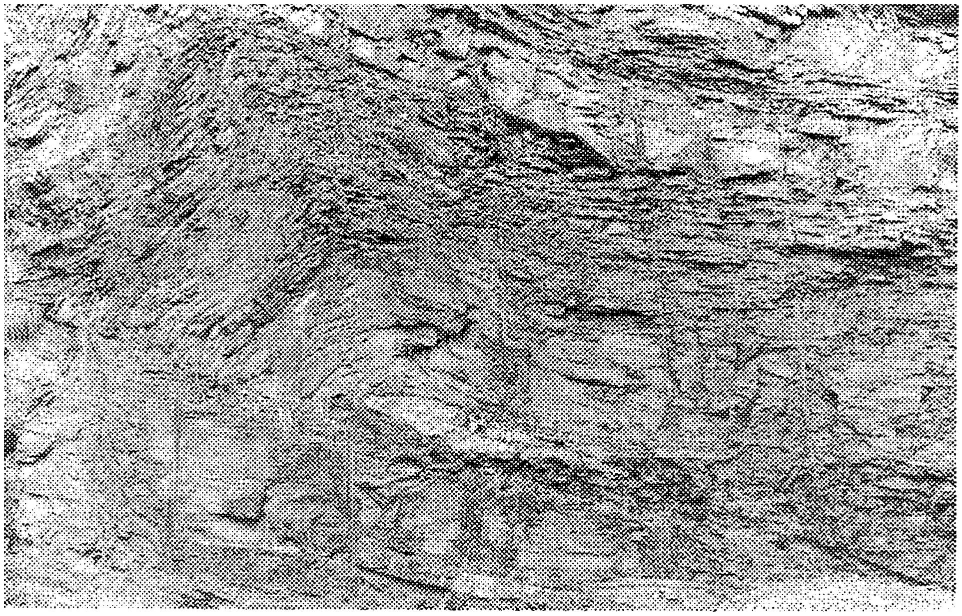
Four main types of fold-accommodation faults are commonly found. Out-of-syncline and into-anticline thrusts are formed because of increase in bed curvature within fold cores, with differential layer-parallel strain also contributing to thrust slip. Depending on the kinematic evolution of the major fold, the thrusts may propagate along the steep or gentle limbs of asymmetric folds or along the hinges of symmetric folds.

Wedge thrusts are primarily formed in competent units because of variations in layer-parallel strain between adjacent units. Hinge wedges occur as multiple nested faults that tend to thicken bedding near fold hinges, whereas limb wedges occur as hanging-wall and/or footwall fault-bend and fault-tip folds, accommodating differences in layer-parallel strain.

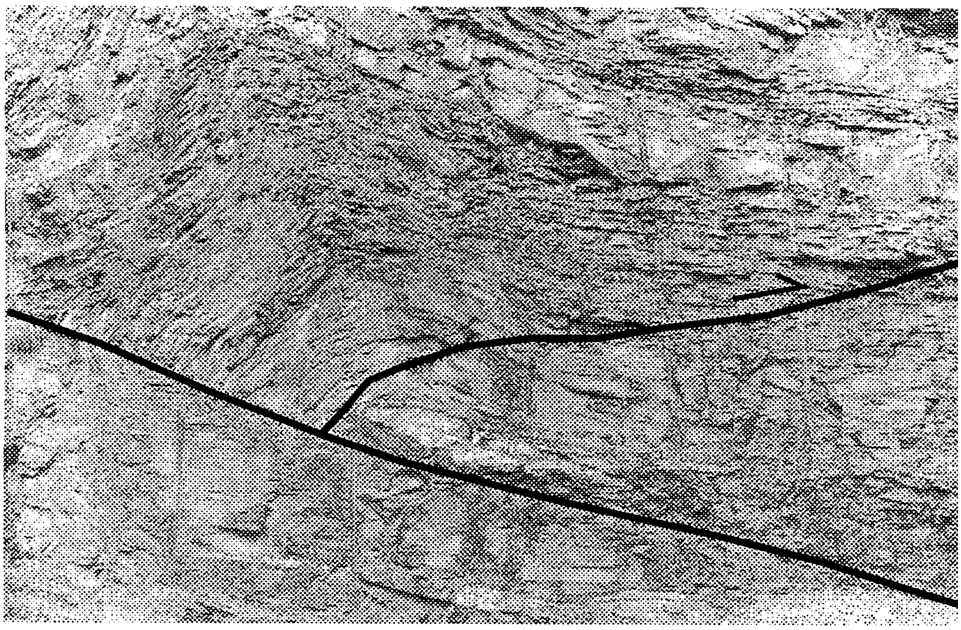
Forelimb thrusts form by a variety of mechanisms. Forelimb space-accommodation thrusts are low-displacement thrusts that resolve strain discontinuities resulting from increased curvature in fold cores. They may link with back thrusts originating in synclines to form forelimb-backlimb thrusts. Forelimb shear thrusts form in the late stages of folding because of forward shear thrusts on the steep forelimbs of folds. A wide range of structures ranging from mesoscopic and microscopic conjugate faults to large-scale single faults may occur.

Back thrusts accommodate hanging-wall strain during the formation of fault-related folds. They may

Figure 22. (a) Back thrust associated with a fault-propagation fold through the Dunlap section in the Appalachian Plateau in Tennessee. The faults are highlighted in (b). The back thrust is rotated on the front limb of the fold and by the synclinal axis on the backlimb.



(a)



(b)

form selectively in more competent units, partially relieving strain associated with rotation of through fault bends. In homogeneous thin-bedded units, characterized by fault-propagation folding, the propagation of the back thrust may keep pace with the rate of propagation of the main thrust. In this case, the fault may also be folded with progressive deformation.

Although all fold-accommodation faults are secondary to major folds, they may involve significant slip and result in the formation of secondary folds of significant scale. The faults may define the geometry and size of hydrocarbon traps. Furthermore, slip distributions on the faults may define the nature of juxtaposition of footwall and hanging-wall units, thereby controlling their sealing capacity. Accurate mapping of these structures is therefore critical in interpreting structural trap geometries and understanding the anatomy of fold and thrust belts.

REFERENCES CITED

Almendinger, R., 1998, Inverse and forward numerical modeling of trishear fault propagation folds: Tectonics, v. 17, p. 640–656.

Barber, D. W., and G. M. Sowers, 1974, A photoelastic study of the effects of surface geometry on fault movements, *in* Advances in rock mechanics: Proceedings of the Third Congress of the International Society of Rock Mechanics, v. 2a, p. 585–590.

Boyer, S. E., 1986, Styles of folding within thrust sheets: examples from the Appalachian and Rocky Mountains of the U.S.A. and Canada: Journal of Structural Geology, v. 8, p. 325–339.

Boyer, S. E., and D. Elliott, 1982, Thrust systems: AAPG Bulletin, v. 66, p. 1196–1230.

Buxtorf, A., 1916, Prognosen und befunde beim Hauenstembasis- und Grenchenberg tunnel und die Bedeutung der letzeren fur die geologie des Juragebirges: Verhandlungen der Naturforschenden Gesellschaft in Basel, v. 27, p. 184–205.

Chester, J. S., and F. M. Chester, 1990, Fault-propagation folds above thrusts with constant dip: Journal of Structural Geology, v. 12, p. 903–910.

Cloos, E., 1964, Wedging, bedding plane slip, and gravity tectonics in the Appalachians (abs.): Geological Society of America Abstracts, v. 76, p. 239.

Colman-Sadd, S. P., 1978, Fold development in Zagros simply folded belt, southwest Iran: AAPG Bulletin, v. 62, p. 984–1003.

Dahlstrom, C. D. A., 1970, Structural geology in the eastern margin of the Canadian Rocky Mountains: Bulletin of Canadian Petroleum Geology, v. 18, p. 332–406.

Dahlstrom, C. D. A., 1990, Geometric constraints derived from the law of conservation of volume and applied to evolutionary models of detachment folding: AAPG Bulletin, v. 74, p. 336–344.

Davis, T. L., J. S. Namson, and S. Gordon, 1996, Structure and hydrocarbon exploration in the transpressive basins of southern California, *in* P. L. Abbott and J. D. Cooper, eds., Field conference guide: Pacific Section AAPG Guidebook 73, Pacific Section SEPM Book 80, p. 189–238.

De Sitter, L. U., 1964, Structural geology, 2d ed.: New York, McGraw-Hill, 551 p.

Douglas, R. J. W., 1958, Mount Head map area, Alberta: Geological Survey of Canada Memoir 291, 241 p.

Elliott, D., 1977, Some aspects of the geometry and mechanics of thrust belts: parts 1 and 2: 8th Annual Seminar of the Canadian Society of Petroleum Geology, unpaginated.

Erslev, E. A., 1991, Trishear fault propagation folding: Geology, v. 19, p. 617–620.

Erslev, E. A., and K. R. Mayborn, 1997, Multiple geometries and modes of fault-propagation folding in the Canadian thrust belt: Journal of Structural Geology, v. 19, p. 321–335.

Faill, R. T., 1969, Kink band structures in the Valley and Ridge province, central Pennsylvania: Geological Society of America Bulletin, v. 80, p. 2539–2550.

Faill, R. T., and R. B. Wells, 1974, Geology and mineral resources of the Millerstown quadrangle, Perry, Juniata, and Snyder counties, Atlas 136: Harrisburg, Commonwealth of Pennsylvania Department of Environmental Resources, 276 p.

Faill, R. T., R. B. Wells, R. P. Nickelsen, and D. M. Hoskins, 1973, Structure and Silurian–Devonian stratigraphy of the Valley and Ridge province, central Pennsylvania, *in* Guidebook for the 38th annual field conference of Pennsylvania geologists: Harrisburg, Department of Environmental Resources, Pennsylvania Geological Survey, 168 p.

Gallup, W. B., 1951, Geology of the Turner Valley oil and gas field, Alberta, Canada, Western Canada sedimentary basin: AAPG Bulletin, v. 35, p. 397–414.

Hardy, S., and M. Ford, 1997, Numerical modeling of trishear fault-propagation folding and associated growth strata: Tectonics, v. 16, p. 841–854.

Heim, A., 1919, Geologie der Schweiz: Leipzig, C. H. Tanchnitz, 704 p.

Hoskins, D. M., 1976, Geology and mineral resources of the Millersburg 15-minute quadrangle, Dauphin, Juniata, Northumberland, Perry, and Snyder counties, Pennsylvania, Atlas 146: Harrisburg, Commonwealth of Pennsylvania Department of Environmental Resources, 38 p.

Huftile, G. J., and R. S. Yeats, 1995, Convergence rates across a displacement transfer zone in the western Transverse Ranges, Ventura basin, California: Journal of Geophysical Research, v. 100, p. 2043–2067.

Jamison, W. J., 1987, Geometric analysis of fold development in overthrust terranes: Journal of Structural Geology, v. 9, p. 207–219.

Medd, D. M., 1996, Triangle zone deformation at the leading edge of the Papuan fold belt, *in* P. G. Buchanan, ed., Petroleum exploration, development and production in Papua New Guinea: Proceedings of the Third Papua New Guinea Petroleum Convention, p. 217–229.

Mitra, S., 1986, Duplex structures and imbricate thrust systems, geometry, structural position and hydrocarbon potential: AAPG Bulletin, v. 70, p. 1087–1112.

Mitra, S., 1987, Regional variations in deformation mechanisms and structural styles in the central Appalachian orogenic belt: Geological Society of America Bulletin, v. 98, p. 569–590.

Mitra, S. 1990, Fault-propagation folds: geometry, kinematic evolution, and hydrocarbon traps: AAPG Bulletin, v. 74, p. 921–945.

Mitra, S., 1992, Balanced structural interpretations in fold and thrust belts, *in* S. Mitra and G. W. Fisher, eds., Structural geology of fold and thrust belts: Baltimore, Johns Hopkins University Press, p. 53–77.

Mitra, S. and J. S. Namson, 1989, Equal-area balancing: American Journal of Science, v. 289, p. 563–599.

Perry, W. J., 1975, Tectonics of the western Valley and Ridge foldbelt, Pendleton County, West Virginia, a summary report: U.S. Geological Survey Journal of Research, v. 3, p. 583–588.

Perry, W. J., 1978, Sequential deformation in the central Appalachians: American Journal Science, v. 278, p. 518–542.

Price, R. A., 1965, Flathead map area, British Columbia: Geological Society of Canada Memoir 336, 221 p.

Serra, S., 1977, Styles of deformation in the ramp regions of overthrust faults: Wyoming Geological Association Handbook, 29th Annual Field Conference, p. 487–498.

Suppe, J., 1985, Principles of structural geology: Englewood Cliffs, New Jersey, Prentice-Hall, 537 p.

Suppe, J., and D. A. Medwedeff, 1984, Fault-propagation folding (abs.): Geological Society of America Abstracts with Programs, v. 16, p. 670.

Suppe, J., and D. A. Medwedeff, 1990, Geometry and kinematics of fault-propagation folding: Eclogae Geologicae Helvetiae, v. 83, p. 409–454.

Willis, B., and R. Willis, 1934, Geologic structures: New York, McGraw Hill, 544 p.

Woodward, N. B., S. E. Boyer, and J. Suppe, 1985, An outline of balanced cross sections: University of Tennessee Department of Geological Sciences Studies in Geology 11, 2d ed., 170 p.

Yeats, R. S., 1983, Large-scale Quaternary detachments in Ventura basin, southern California: Journal of Geophysical Research, v. 88, p. 569–583.