

THIN-SKINNED DEFORMATION OVER SALT

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I. INTRODUCTION - SALT STRUCTURES AND FOLD BELTS

There are more than a dozen fold-and-thrust belts that have at some time, and at least in places, moved atop a layer of evaporites, usually including salt. The Triassic evaporites beneath both the Jura (Laubscher, 1972; Gwinner, 1978) and Pyrenees (Brinkman and Logters, 1968; Liechti, 1968) are intimately involved in the deformation resulting from horizontal compression. The same is true of the Cambrian Saline River formation beneath the Franklin Mountains in northwestern Canada (e.g., Cook and Aitken, 1973; Aitken et al., 1982) and the Ordovician Bay Fiord and Otto Fiord formations beneath the Canadian Arctic fold belts (e.g. Price and Douglas, 1972; Fox, 1984). The Cambrian salt beneath the Salt Range of Pakistan provides a weak detachment upon which thin-skinned deformation can advance far beyond that in adjacent, apparently salt-less, areas along strike (e.g., Sarwar and DeJong, 1979; Seeber et al., 1981; Burbank, 1983). The tectonic development of the Zagros belt is dominated by the horizontal decoupling and vertical diapirism of massive salt beds of Late Precambrian and Miocene age (e.g., Farhoudi, 1978; Colman-Sadd, 1978). Other mountain belts whose tectonics are strongly influenced by evaporites include the

Such deformation of the overlying rocks will tend to have a geometry controlled by the regional state of stress, and that geometry will be inherited by the resultant diapiric structures, whether of a piercement or non-piercement type. The very close correspondence between pre-existing faults and the elongated salt structures of the southern North Sea and northern Germany is a particularly well known example of this relationship.

There are numerous examples of prominent elongate salt-related structures associated with fold-and-thrust belts (for example, see Figure 1) where salt is intruded along thrust faults or at anticlines (e.g., Harding and Lowell, 1979). In addition to the Zagros (e.g., Stocklin and Nabavi, 1973), these include the Carpathian (e.g., Paraschiv and Olteanu, 1970), Sierra Madre Oriental (e.g., Weide and Martinez, 1970), Pyrenees (e.g., Liechti, 1968), Amadeus Basin (e.g., McNaughton et al., 1968) and Atlas (e.g., Tortochaux, 1978) fold belts.

II. IDEAS ABOUT THIN-SKINNED DEFORMATION

A. Historical development

Over a century ago it was recognized that the Glarus Thrust in Switzerland represented a major overthrust (Heim, 1871). Permian rocks in the upper plate of the thrust had "traveled" many km north by northwest over the Mesozoic rocks of the lower plate. This observation sparked a debate among Alpine geologists that was to last many decades: were these rocks pushed from the rear or did they slide down hill? It was soon recognized that frictional resistance along the basal decollement and the failure strength of rock were two major parameters controlling the length of overthrust sheets. By early in this century, the brittle strength of rocks was well documented in the laboratory. Smoluchowski (1909) recognized that for an overthrust to move over a horizontal surface, it would have to be pushed from the side with a force equal to the total frictional resistance along

Sierra Madre Oriental of Mexico (Rogers et al., 1962; de Cserna, 1971), the Cordillera Oriental of Colombia (Campbell and Burgl, 1965; McLaughlin, 1972), the Atlas Mountains of Tunisia and Algeria (Tortochaux, 1978), the southern Urals (Nalivkin, 1973; Kazantsev and Kamaletdinov, 1977), the Tadjik fold-and-thrust belt (Leith, 1984), and the anticlinal province of the Amadeus Basin in Australia (McNaughton et al., 1968).

"Thin-skinned" mountain belts result from horizontal compression (e.g. Chapple, 1978). Their structures are predominantly oriented parallel to the intermediate stress axis (Anderson, 1942), in the horizontal direction perpendicular to the maximum shortening. Deformation of the overlying rocks can produce both the overburden inhomogeneity and stress instability required to initiate diapiric growth, and the geometric ordering required to initiate systematically elongate salt structures. Even where there is no development of salt domes, such as in the fold belts of the Appalachian Plateau (e.g. Gwinn, 1964) and the Jura (e.g., Laubscher, 1977), the folds contain considerable anticlinal thickening of the salt, associated with synclinal thinning of salt.

Salt diapirism is usually associated with faulting in overlying rock. The growth of salt domes requires a density inversion and leads to faulting of the displaced younger strata (e.g. Nettleton, 1934). In the absence of significant horizontal differential stresses, such faulting and the associated salt domes, should be expected to have no strongly preferred orientation in map-view.

Conversely, deformation of the overlying sediments and the tectonic forces that cause it can be major factors in controlling the development of salt diapirs. When folding and faulting alters the thickness of the overlying section or the depth to the top of the potentially mobile evaporite, then it enhances the mechanical instability required to initiate diapiric movement.

which tapers toward the margin of the mountain belt. 4) the basal layer beneath the fold belt commonly consists of rock that is, for some reason, relatively weak (e.g. commonly a shale).

The interiors of mountain belts are in fact deformed in compression, as was predicted by Smoluchowski (1909) and Hubbert and Rubey (1959). Thus, a taper is produced in the direction of the craton, with a topographic surface that slopes gently toward the craton and a decollement surface that slopes away from the craton (Figure 3). Chapple (1978) showed that a perfectly plastic fold-and-thrust belt over a very weak basal layer can overthrust its base if there is sufficient topographic slope.

Davis et al. (1983) presented a simple model for a cohesionless, time-independent, homogeneous and isotropic Coulomb wedge that deforms in a manner analogous to soil being pushed by a bulldozer. For most fold-and-thrust belts, this model does not require an extremely weak basal layer. Despite its simplicity, this model proved successful in predicting the gross geometry of the active fold-and-thrust belt of western Taiwan, particularly when the role of cohesion was taken into account (Dahlen et al., 1984). This wedge deforms until it attains a steady state or critical taper and then slides stably, continuing to grow self-similarly as additional material is accreted at the toe (Davis et al., 1983; Dahlen et al., 1984). The thick end of a fold-and-thrust belt wedge is at the core of the mountain range. In the



Figure 3. A cross-section through the southern Appalachians (Poeder et al., 1978) modified after Davis et al. (1983). The intense folding and thrusting is concentrated in the wedge-shaped overthrust belt, which overlies a layer of decollement. Note that structures verge predominantly toward the craton.

Hubbert-Rubey pore-fluid pressure coefficient, is given by the relation

$$\lambda = P_p / \rho g z \quad (1b)$$

Here, P_p is the pore fluid pressure, ρ is the mean density of the sedimentary overburden, g is the gravitational acceleration, and z is the depth. The pore fluid pressure is termed 'lithostatic' ($\lambda=1$) if it is equal to the total overburden pressure of the overlying sediments. With unimpeded pore space connectivity to the water table (assumed to be at the surface), the pore fluid pressure is termed 'hydrostatic'. In that case $\lambda = (\rho_w / \rho) \approx 0.4$, where ρ_w is the density of water.

Data from laboratory experiments enabled the calculation of the size and length of an overthrust sheet that could be pushed from the rear with elevated pore pressures along its base. Assuming reasonable rock strengths, Hubbert and Rubey (1959) showed that known overthrusts could have been pushed horizontally from the rear if sliding was facilitated by nearly lithostatic pore-fluid pressures.

B. Critical wedge theory

Chapple (1978) listed several characteristics that are shared by foreland fold-and-thrust belts and that are essential to consider in modeling their mechanics. These are: 1) the basal surface of detachment or decollement, below which there is little deformation, dips toward the interior of the mountain belt. This dip direction is usually away from the craton. In other words, most fold-and-thrust belts were pushed up hill and did not benefit from the help of gravity gliding. 2) a large horizontal compression occurs in the rock above the basal flattening in response to this large compression. 3) in cross section, the deformed rock mass has a characteristic wedge shape

B. Mechanisms for reduction of friction

There are at least two common classes of thin-skinned deformation in which overthrusting is met with very weak resistance to sliding. The first of these classes consists of thin-skinned wedges with very high pore-fluid pressures. Although near-lithostatic pore pressures in fold-and-thrust belts appear to be very rare (e.g., Fertl, 1976; Suppe and Wittke, 1977), there are many accretionary prisms in which the accreting sediments are very porous and must dewater as part of the diagenetic process (e.g., von Huene and Lee, 1982; Moore and Biju-Duval, 1984). There is a growing body of data suggesting that deformation at the frontal toe of at least some accretionary prisms takes place under conditions of extremely high pore pressures. As demonstrated by Hubbert and Rubey (1959), this permits overthrusting (or, in this case, subduction) at very low shear stresses.

The other special class of thin-skinned deformation includes those fold-and-thrust belts that overthrust, at least in part, a detachment zone in an evaporite, and it is that class with which this paper is concerned. There are more than a dozen fold-and-thrust belts that have formed atop an evaporitic layer.

For moderate geothermal gradients (15° to $25^{\circ}\text{C km}^{-1}$), a temperature of 100°C is found at 3 to 5 km below the surface. Let us assume that at this depth, there is a detachment along a 100 m thick salt layer, where 1 cm/yr of slip takes place.

Equation (3) indicates that at this strain rate ($\approx 3.10^{-12}\text{sec}^{-1}$) the differential stress required to drive this slip is only 1 MPa, which corresponds to a maximum shear stress of 500 kPa (5 bars). Varying the strain rate by two orders of magnitude changes this result by factor of less than 3, with higher strain rates requiring higher differential stresses.

Evaporites in general (and rock salt in particular) are much weaker than any other common rock type, including shales.

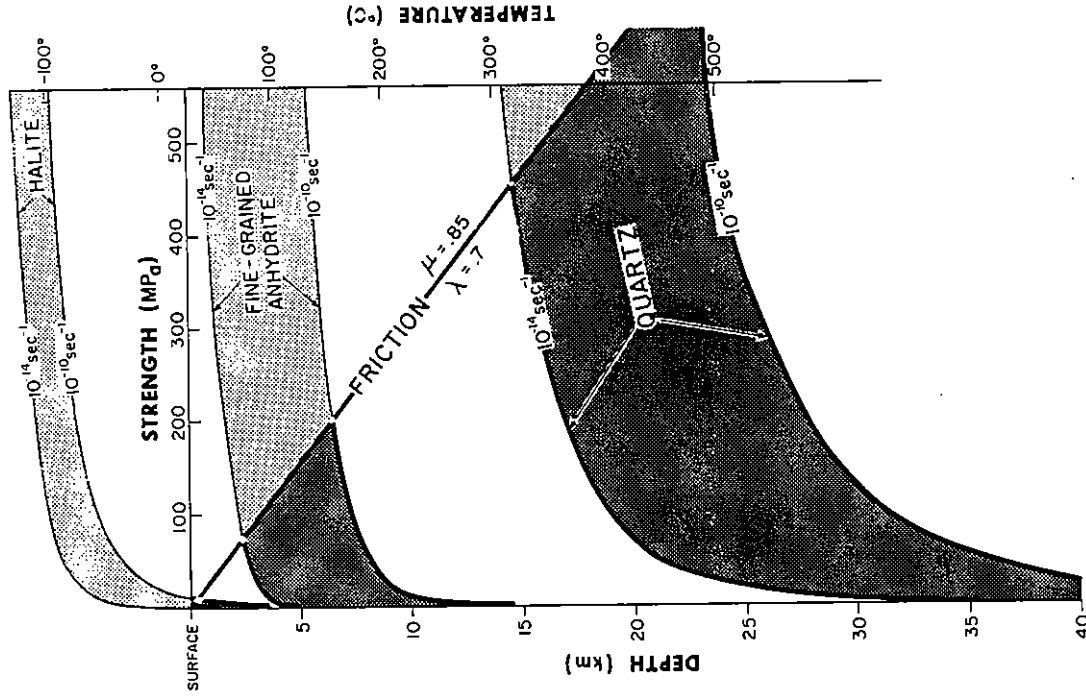


Figure 4. Strengths of common rock types versus depth and temperature, showing the relative weakness of halite. The frictional strength assumes that the coefficient of friction $\mu=0.85$ (Byerlee, 1978), density = 2.75 g/cc , and pore fluid pressure ratio $\lambda=0.7$. The flow strengths are as follows: quartz from Brace and Kohlstedt (1980), halite from Carter and Hansen, (1983), and anhydrite from Muller et al., (1981).

foldbelt that is longer as well as more narrowly tapered. Thus, the part of a foldbelt that overthrusts a salt basin should be broader in map-view and more narrowly tapered in cross-section over the salt than at points along strike that are not over the salt. As for all of the arguments made in this paper, the same should be true, though to a somewhat lesser degree, for overthrusting on anhydrite, which has a brittle-ductile transition somewhat deeper than that of halite (Figure 4).

2. Stress orientations and vergence

Another parameter controlled by the presence of a very weak basal detachment is the dip ψ_b at which the axis of maximum compressive stress σ_1 dips toward the foreland with respect to the basal detachment. The orientation of the principal stress axes is important because of the following relation to deformation. The Coulomb failure criterion is satisfied most readily along the two planes containing the σ_2 (intermediate) stress axis and inclined about the σ_1 axis at an angle θ defined by the simple relation

$$\theta = \pm (45^\circ - \phi/2) . \quad (5)$$

In order to have the proper sense of shear traction along the basal decollement, the σ_1 (maximum compression) axis must dip toward the foreland, as shown by Hafner (1951) for a rectangularly shaped overthrust. The angle ψ_b at which the σ_1 axis dips with respect to the basal decollement for a thrust wedge has been calculated by Davis et al. (1983) and Dahlen et al. (1984). The cohesive strength can be neglected in this calculation if it is much smaller than the component of strength due to rock sliding friction, such that $S_0 \ll \rho g z \mu (1-\lambda)$. Here, S_0 , ρ , g , z , μ , and λ are the cohesive strength, rock density, gravitational acceleration, depth, rock friction coefficient, and Hubbert-Rubey pore-fluid pressure coefficient, respectively. In the limit of negligible cohesion

$$\psi_b = \frac{1}{2} \arcsin(\sin\phi_b/\sin\phi) - \phi_b/2 , \quad (6a)$$

where ϕ and ϕ_b are the internal friction angle of the wedge and the sliding friction angle for its base, respectively. This relation can be demonstrated by using a Mohr-Coulomb diagram (Davis et al., 1983). Dahlen et al. (1984) show that in the presence of significant wedge cohesion, it is necessary to solve iteratively for ψ_b using the relation

$$[\mu(1-\lambda) + (S_0/\rho g)] \sin 2\psi_b = \mu_b (1-\lambda) [(1+\mu_b)^2 - \mu_b \cos 2\psi_b] \quad (6b)$$

where $\mu_b \tan \phi_b$ is the coefficient of sliding friction along the basal detachment layer. The magnitude of ψ_b at the base of the deforming wedge is very strongly dependent upon shear traction which can be supported across the basal detachment. Dahlen et al. (1984) have calculated that beneath the toe of the Taiwan fold-and-thrust belt, $\psi_b = 12^\circ$. If the coupling were weakened by replacing the Talu Shale with a thick layer of salt, then the basal friction μ_b would be replaced by $\tau_0/\rho g z$, where τ_0 is a normal-stress independent strength, as for the critical taper calculation in Section III.C.1. In that case, with $\tau_0 = 1$ MPa or less, then ψ_b would be considerably less than 1° .

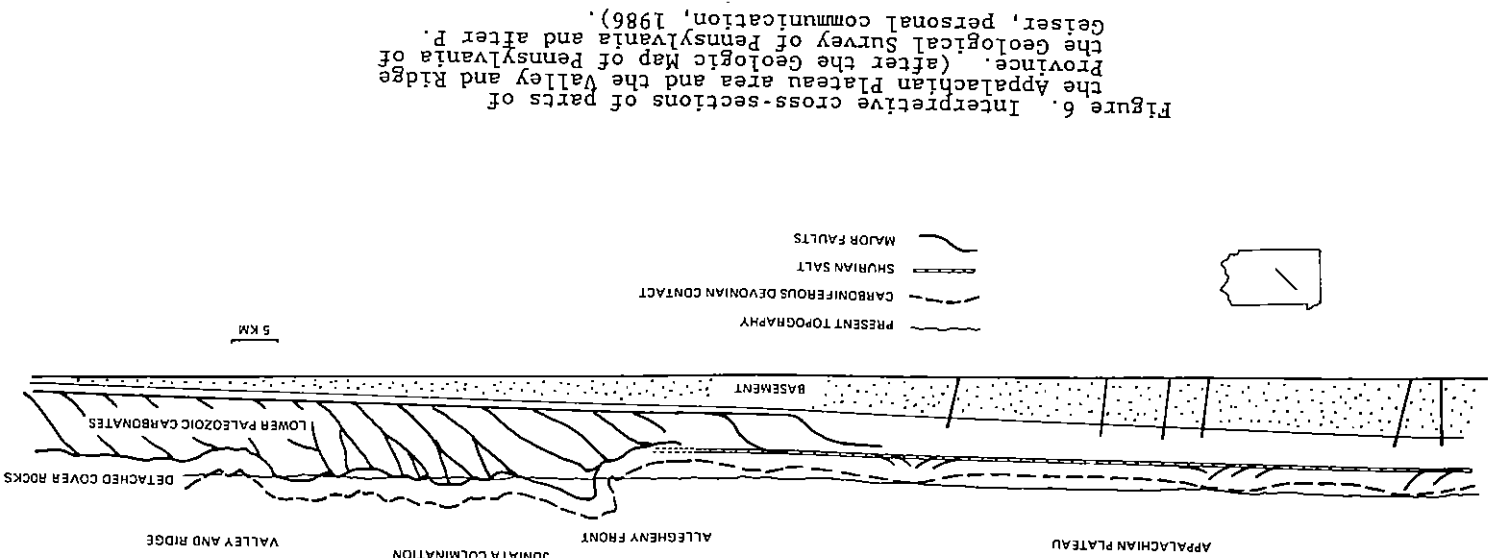
Both sets of possible slip planes are symmetric about the σ_1 axis, which dips forward at an angle ψ_b . Therefore (Figure 5) the forward verging slip planes have an optimal dip at an angle $\delta_f = \theta - \psi_b = 45^\circ - (\phi/2) - \psi_b$ and the backward vergent ones dip at $\delta_b = \phi + \psi_b = 45^\circ - (\phi/2) + \psi_b$. The fact that the forward verging slip planes dip more shallowly (by a margin of $2\psi_b$) probably explains why they are the more common of the two in most thrust belts (Davis and Engelder, 1985). The reason for this is simple. The shallower dip of the forward verging slip planes permits a greater amount of horizontal shortening for the same increase in gravitational potential energy. Forward verging thrusts are also

the Plateau (Roggers, 1963; Wiltchko and Chapple, 1977). However, the styles of folding are quite different. Only one Plateau fold has structural relief at the surface exceeding 800 m or a flank dip greater than 10° , but the first fold in the Valley and Ridge Province has relief of 8 km and is overturned (Roggers, 1963). In addition, all Valley and Ridge folds (Figure 6b) are much steeper on their northwestern sides, but Plateau folds are much less asymmetric, with only a slight preference toward a greater steepness on the southeastern side (Figure 6a). This tendency toward slightly oversteepened southwest limbs is most prominent in the northwestern part of the Plateau, where amplitudes are the smallest (Sherrill, 1934; Gwinn, 1964). The contrast between the two provinces extends to faulting as well. Although there are many surface thrust faults (all dipping to the southeast and verging to the northwest) in the Valley and Ridge, there are extremely few surface faults in the Plateau, and those that have been found do not seem to have any marked preference in vergence (Roggers, 1963).

This lack of a strongly preferred direction of structural vergence is a very common attribute of fold-and-thrust belts over salt. For example, both the folds and the thrust faults of the Franklin Mountains are inconsistent in their direction of asymmetry. Indeed, there are several places in the northern Franklin Mountains (Figure 7) where thrust faults with opposite senses of transport occur near to, and along strike from, each other (Cook and Aitken, 1973).

B. Fold geometry

A cross section through the Appalachian Plateau (Figure 6a) shows that Coulomb-like failure is localized in some of the more rigid formations (sandstone and limestone) just above the salt decollement, in the core of widely spaced anticlines. Engelder and Geiser (1979) show that above those rigid layers the



but not in it. Furthermore, the higher normal stresses of the thickened section repress additional Coulomb-like failure in that area.

The location of additional Coulomb failure is a function of frictional dissipation of the traction pushing the Appalachian Plateau toward the foreland. This dissipation is low. But, with everything else equal, the traction on the interior side of the thickened section is larger than the traction on the exterior (cratonward side).

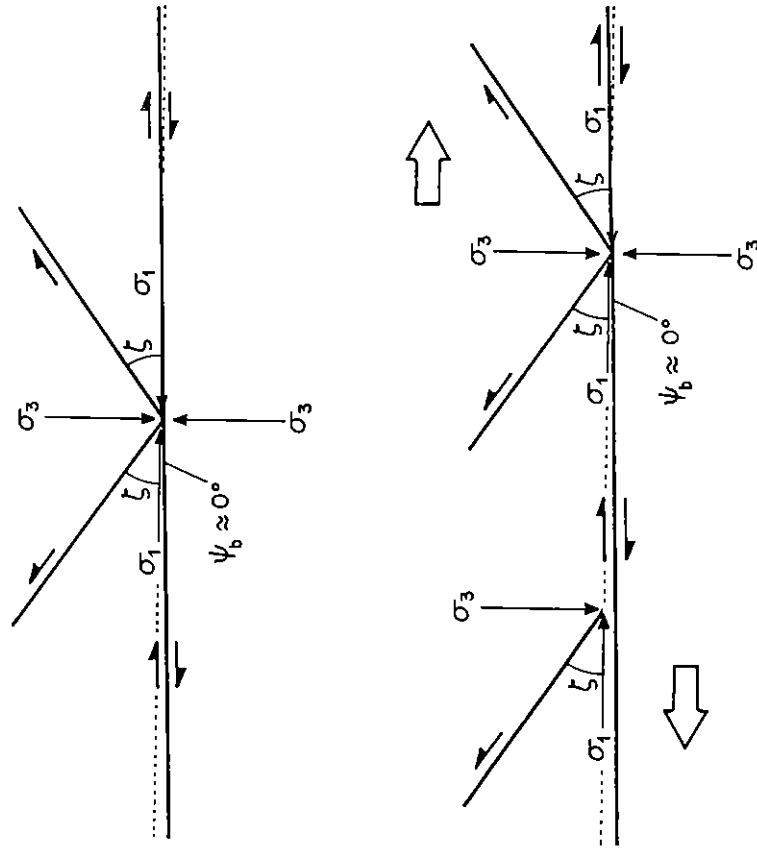


Figure 9. Proposed mechanism for the subtle asymmetry of Appalachian Plateau folds, with the steeper limb facing toward the interior of the mountain belt.
 (a) Initial, symmetric deformation.
 (b) Development of subsequent additional backward vergent thrust(s) in response to greater traction immediately behind the original thrusts.

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Failure is then favored on the interior side of the thickened section by virtue of a slightly larger traction there.

For Coulomb failure on the interior side of the thickened section, failure along back thrusts is favored where the fracture propagates back into a thinner section. Here the normal stress across the failure planes for back thrusts is lower than normal stress for forward thrusts. This gives the section a second back thrust and further steepens the asymmetry of the folds on the Appalachian Plateau (Figure 9b). Several asymmetric folds have apparently developed across the Appalachian Plateau by this process.

Plateau folds within 50 km of the structural front with the Valley and Ridge Province have structural relief close to the local thickness of the salt. Wiltchko and Chapple (1977) suggest that this relief is produced as a result of synclinal thinning of salt. Near the structural front the synclines appear to be essentially depleted, but further away from the structural front the synclines are relatively less depleted. This is taken to be an explanation for the reduction in structural relief of the folds towards the northwest. In some areas, anhydrite apparently takes part in the flowage along with the halite. Wiltchko and Chapple (1977) have also pointed out that as a syncline becomes depleted of salt, it must become relatively more difficult to continue the thinning of the synclinal salt because the stress required for a constant rate of thinning is proportional to the inverse cube of the layer thickness.

The strength of the basal decollement in the salt layer also increases with synclinal thinning. A thinner salt layer must deform at a proportionally higher strain rate in order to attain a given rate of overthrusting slip. However, as we noted in Section III.B, the shear stress is relatively insensitive to the strain rate (eqn. (3)). Nevertheless, forward oversteepening of folds in the Zagros fold belt, where the Cambrian salt is

thickening by widespread Coulomb failure. This would explain the lack of major thrust faults everywhere in the Plateau except occasionally just above the salt and the predominance of the observed homogeneous, lower stress, lower strain rate deformation mechanism, layer-parallel shortening.

D. Relation of folding salt basin

There is a very close relation between Appalachian Plateau deformation and the distribution of the Salina salt, as is clear from the distribution of layer-parallel shortening (Figure 10). The transition from Valley and Ridge folding and faulting to Appalachian Plateau folding occurs at the southeastern edge of the salt basin, where the decollement steps up section until it reaches the mechanically favored weak Silurian salt (Figure 6). A very similar transfer of slip is seen to occur in other fold belts that encroach upon a salt basin for part of their lengths.

One particularly interesting example of this process is found in the Franklin Mountains of northwestern Canada. Where there are no Saline River Formation evaporites, the part of the section younger than Late Cambrian is effectively pinned to basement. However, where the frontal thrusts of the Mackenzie Arc reach part of the basin containing the salt, there is a transfer of slip to that level. There, deformation does not stop with the frontal Mackenzie thrust fault, but continues another 100 km or so onto the craton (Figure 9). The front of significant folding does not reach the Peel Plain even though there is salt beneath it, because there is no salt to the southwest along the Mackenzie front into which slip could be transferred (Aitken and Cook, 1975; Aitken et al., 1982).

The Burning Springs Anticline at the southwestern edge of the Appalachian Plateau is an anomalously trending, large amplitude structure. This is not a normal Plateau fold. It appears to be related to a transfer of slip to the surface where there was no longer sufficient salt in which to form a weak

decollement (Rodgers, 1963). The Burning Springs Anticline follows the trend of the edge of the salt, and might be likened to the piling up of a rug adjacent to where it is pinned down.

V. SALT-RELATED STRUCTURES IN OTHER FOLDBELTS

A. Box Folds

The folds of the Appalachian Plateau (Figures 6,8) are typically characterized by splay faults off the basal decollement which produce salt-cored anticlinal uplift (Gwinn, 1964). These folds are much like the drape folds described by Stearns (1971). Salt has thickened under these regularly spaced anticlines (Wiltchko and Chapple, 1977). Another style of folding observed above salt is that found in the Jura, where non-sinusoidal fold: in a considerably thinner overthrusting section have been described as a superposition of different kinds of instabilities: in space and time (Laubscher, 1972, 1977). They are most realistically modeled as box fold structures with very large (hundreds of meters across) kink bands with rounded edges (Figure 11).

It is possible that the presence of either Appalachian Plateau or Jura type folding may depend upon whether there is a major 'rigid' structural member low in the section as is the case in the Appalachians (the Onondaga Limestone and Oriskany Sandstone). The difference between fold style in the Appalachian Plateau and the Jura may also be related to differences in the thickness and the mechanical homogeneity of the sections, with the Jura comparatively thin and homogeneous.

B. Effects of basement structure

Basement faults and warps are known to influence deformation in many fold-and-thrust belts. Wiltchko and Eastman (1983) have demonstrated a number of mechanisms for the generation of stress concentrations due to basement structure. They point out that

show the strong structural discrimination between their forward and backward directions typically found in strong-basal mountain belts.

C. Effects of the distribution of salt

Oroclines are curved mountain belts, which are usually assumed to have started out straight. Because salt facilitates overthrusting, the presence of a salt basin provides an obvious thin-skinned mechanism for the development of oroclines. However, oroclines can form in other ways as well. The original shape of a continental margin is likely to be ziz-zag in form because of the effect of transform faulting. It has been suggested (Thomas, 1977) that margin re-entrants become structural salients and margin promontories become structural recesses. Pre-existing basement structure is the obvious explanation for the Mackenzie Arc of northwestern Canada, but it cannot explain the outward extension of the Franklin Mountains and the Colville Hills (Figure 7), which is a thin-skinned phenomenon related to the distribution of the Cambrian salt.

The distribution of thin-skinned thrusting is one of the many aspects of Andean tectonics that appears to be very closely related to the angle of subduction of the Nazca plate (e.g., Jordan et al., 1983). However, Jensen (1984) has pointed out that there is a nearly one-to-one correspondence between the distribution of thin-skinned thrusting in Chile and Argentina and that of Oxfordian (Upper Jurassic) gypsum.

As the deformation in an evaporite-basal foldbelt moves far ahead of the position of the structural front in areas where there is no salt, there are inevitably going to be drag and rotational features at the edges of the salt layer along strike. The Burning Springs Anticline of the Appalachian Plateau (Section III) and the wrapping of folds around the Coahuila Peninsula in the Sierra Madre Oriental (Kleist, 1984) are two probable examples of this phenomenon. Such rotations, if they exist,

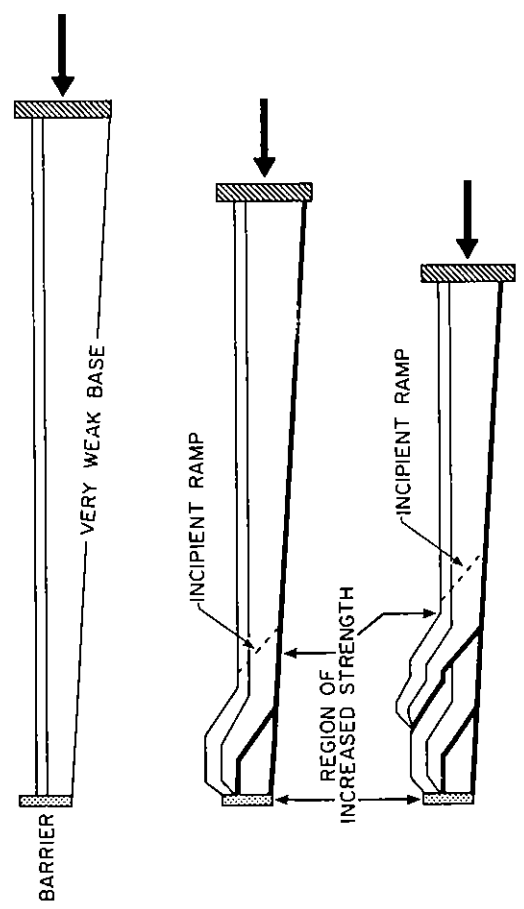


Figure 12. Schematic cross-section through a weak-basal foldbelt with a large basement offset at its front, (such as the Tadjik foldbelt), showing a backward migration of structures (after Leith and Alvarez, 1986).

slide well below its failure strength everywhere except near its northern end where there is a large mechanical barrier (the Illiac fault, which offsets basement by 3 km). This barrier causes localized failure and buildup of topography in much the same way as the depletion of the Salina salt in West Virginia caused the development of the Burning Springs Anticline of the Appalachian Plateau.

In summary, foldbelts over salt can act almost like stress guides, with relatively little shear traction exerted on its bottom (in salt), as well as none exerted on its top (in air). This is the essential reason why they can respond in such a complicated manner to a major basement uplift, as in the Salt Range (e.g., Yeats and Lawrence, 1985) or the Tadjik fold and thrust belt (e.g., Leith, 1984), and why they generally do not

stress axis over a detachment in salt is also consistent with the commonly observed lack of consistent vergence in structures within salt-basal overthrust belts.

VII. ACKNOWLEDGEMENTS

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should be reflected in paleomagnetic declinations. The results to date are mixed. Paleodeclinations from the Salt Range (Opdyke et al., 1982) and Sierra Madre Oriental (Kleist et al., 1984) are consistent with the expected degree of rotation due to the uneven advance of the overthrust, but at least some data from the Jura and the Appalachian Plateau are not (Eldredge et al., 1985). A recent study in the Appalachian Plateau (Kent and Miller, 1986) finds evidence for oroclinal bending and suggests that previous negative results there can be attributed to post-folding remagnetization. However, even if this is the case, such bending is probably more a response to the basement structure of the marginal salient than to along-strike variations in thin-skinned deformation.

The Verkhoyansk fold belt in Siberia is a north-trending Mesozoic foldbelt that separates the flat-lying upper Precambrian and Paleozoic rocks of the Siberian platform from the Paleozoic Chersky foldbelt and Kolyma massif to the east. Permian clastic rocks over 7000 m thick are deformed into large, linear folds and are bordered on the west by a thick section of syn-orogenic Cretaceous sediments derived from the Verkhoyansk foldbelt (Churkin, 1972). The Verkhoyansk complex probably represents the Paleozoic eastern continental margin of the Siberian plate, deformed in the Mesozoic by collision with the North American plate (Fujita, 1978; Fujita and Newberry, 1982). The Verkhoyansk foldbelt matches the characteristics of a "typical" evaporite-based fold-and-thrust belt so well that we consider the existence of undetected evaporites there to be a distinct possibility (Davis and Engelder, 1985). Published data indicate that some salt and a greater amount of other evaporites have been found in the area, but little is known of their extent (Zarkhov, 1981).

VI. SUMMARY

Although there is probably not a very great difference in strength between the basal detachment zone and the overthrusting rocks in many thin-skinned mountain belts, this is not true of overthrusting over salt. The style of deformation in thin-skinned fold-and-thrust belts is critically dependent upon the resistance to sliding along the detachment between the mass of deforming sediments and the underlying rocks. Evaporites can provide an extremely weak horizon within which a basal detachment can form and along which only a relatively small shear traction can be supported. Fold-and-thrust belts that form atop a salt layer share several readily observable characteristics (Davis and Engelder, 1985). The weakness of a detachment in salt permits a fold-and-thrust belt to maintain a much narrower cross-sectional taper (as little as a few tenths of 1°) than is commonly observed in the absence of salt (8° or more). This is reflected in the great width of many salt-basal fold belts, which project far outward toward the craton map view. Along strike, at the boundary between the parts of the fold belt that do and do not ride on salt, this difference in basal resistance often produces complex structures related to the accommodation of the different amounts of overthrusting in the two areas. Rodgers (1963) explained the anomalously trending Burning Springs anticline near the southwestern edge of the Silurian salt beneath the Appalachian Plateau as just this sort of salt-termination differential-drag structure. En echelon folding (e.g. Norris, 1972) and rotation about a salt-less hinge area (e.g., Seeber and Jacob, 1977; Crawford, 1974) appear to be other ways in which overthrusting rocks accommodate a large contrast in resistance along the detachment between areas with and without salt. The predicted subhorizontal orientation of the maximum principal

over the block fault. These sheets now lie flat atop the autochthon to the north. Thus, the crustal structure inherited from a Mesozoic extensional phase has strongly influenced the Cenozoic pattern of deformation, producing a fold-and-thrust belt in which structures are strongly concentrated toward the northern (toe) end, near the basement fault. Apparently, the development of the thrust system has included the progressive overlapping of thrusts, with the later thrusts having formed internal to the older thrusts which they subsequently overrode. This implies that thrusts in front of a basement offset in at least this salt-based fold-and-thrust belt develop in a reverse sequence, away from the craton (Leith and Alvarez, 1986).

Unlike foldbelts without basal evaporites, salt-basal foldbelts are likely to display stress-guide behavior, because they come close to having a zero shear-stress boundary on the bottom (in salt) as well as on top (in air). When confined and compressed (as is the case in the Tadjik foldbelt), a stress guide (the sedimentary pile) will break at its thinnest (and therefore weakest) point. That thinnest point is located at the outer, marginal edge of the basin; in the case of the Tadjik foldbelt that is immediately adjacent to the Illiac fault. Thrusting leads to a buildup of taper at and a strengthening of the toe in front of the basement fault. Eventually, the frontal thrust must lock, leading to fracture of the adjacent sheet. The process continues: displacement of a sheet on a new ramp; buildup of topography and strength at the toe; and formation of a new internal thrust. The result (Figure 12) is the anomalous, backward migrating, thrust sequencing over the basement normal fault at the front of the Tadjik foldbelt (Leith and Alvarez, 1986).

The original depositional geometry of the Tadjik Basin apparently exceeds the negligible critical taper needed for it to overthrust the Jurassic salt. This allows the thrust sheet to

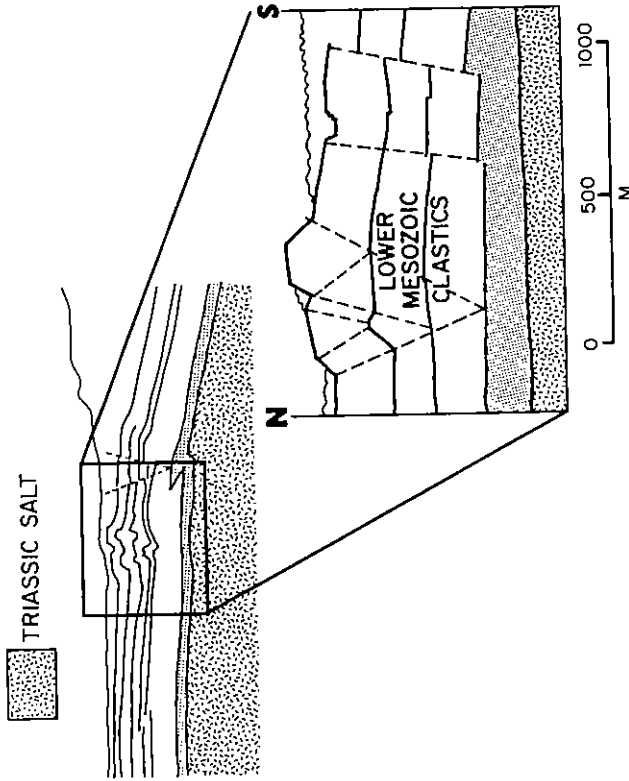


Figure 11. A cross-section through a Jura box fold (after Laubscher, 1977).

small number of stress concentrations, generated by an equally small number of basement structures, can control the gross geometry of structure in an entire fold-and-thrust belt. We suggest that this conclusion may be particularly applicable to salt-basal foldbelts which, as we have seen, commonly lack a consistent preferred direction of structural vergence.

The basal decollement of the Tadjik (Vakhsh) foldbelt in the Soviet Union is in a thick Jurassic salt (Leith et al., 1981). Recent studies have determined that deformation in the Cretaceous and Tertiary strata of the fold-and-thrust belt is, in the central Depression, conspicuously absent north of a buried basement fault that marks both the hinge-zone of the late Mesozoic passive margin and the northern boundary of the evaporite (decollement) horizon. However, in the eastern part of the Depression, thrusting has moved coherent thrust sheets north,

exceptionally mobile, has been attributed to increased basal resistance in areas of synclinal salt depletion (Colman-Sadd, 1978).

C. Strain and Paleostress

The sediments cratonward (to the northwest) of frontal folds in the Appalachian Plateau have undergone layer-parallel shortening by pressure solution. Measurements by Engelder (1979) and by Engelder and Geiser (1979) indicate that the amount of horizontal shortening by pressure solution and intergranular mechanisms is closely related to the distribution of the Silurian salt. Along each of the three traverses in Figure 10, compressive strain increases from very low levels (<2%) where there is no significant Silurian salt to 15% over areas with thick salt. Apparently, in the absence of salt, strata younger than Silurian are strongly coupled to the rocks below and did not deform, but where the Silurian includes significant salt deformation was significant.

The driving stresses of the Alleghanian Orogeny in the Appalachian Plateau have been estimated using several different techniques (e.g., Prucha, 1968; Rutter, 1976; Engelder, 1982). Engelder (1982) inferred that the level of differential stress was 6.2 MPa by observing the axial canals of crinoid columnals within the wedge, which acted as stress concentrators and caused twinning adjacent to the canals. Residual stress measured in rocks from more deeply buried portions of the Appalachian Plateau wedge averages 11 MPa (Engelder and Geiser, 1984). These data indicate that shortening took place at relatively low differential stresses within the wedge, and provide a constraint on the mechanism for the development of the Appalachian Plateau structures over the Silurian salt.

One conclusion that can be drawn from these data is that after the initiation of the initial blind thrusting and splay faults, much of the deformation took place in a more distributed,

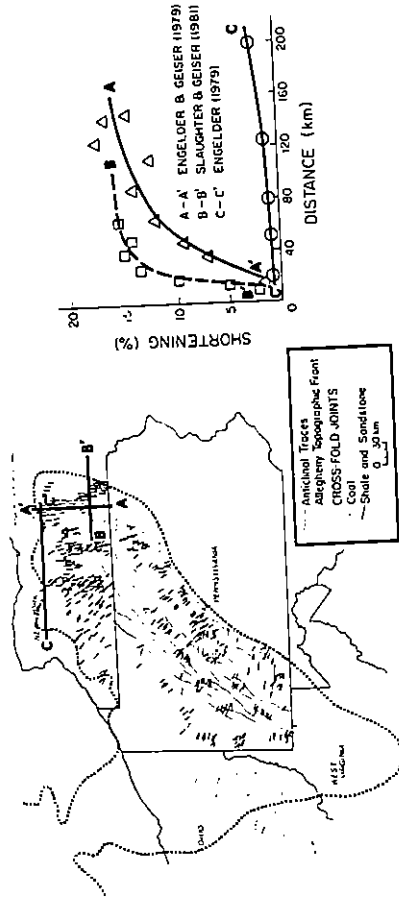


Figure 10. (a) A map of the Appalachian Plateau area, showing the limit of the Silurian salt basin and the locations of three lines, along which measured shortening has been plotted in part (b) (after Davis and Engelder, 1985).

long-term manner, at a relatively low strain rate and a relatively low differential stress. The reasoning is as follows: in the narrowly tapered Plateau, there is no marked deformation front or frontal thrust at which deformation is concentrated. Therefore, at any given time the locus of shortening is quite large (a sizeable fraction of the whole Plateau, as opposed to merely a single frontal thrust). With shortening distributed over a wide area, the strain rate at any given point in space and time must be quite low, apparently low enough to make layer-parallel shortening a viable mechanism for accommodating most of the shortening in the northwestern part of the Plateau.

Given the present taper ($\approx 1.5^\circ$) of the post-Silurian strata (Section 4.A.), it appears that the original, pre-orogenic basinal taper of the Appalachian Plateau may well have been greater than the Coulomb critical taper (eqn. 4) at which high-stress Coulomb failure will take place within the wedge. If the original taper of the Appalachian Plateau wedge exceeded the critical taper, then there would have been no need for further

(Figure 8) is opposite that for most fold-and-thrust belts with forward stepping thrusts. A consequence of this thrusting is that many of the folds of the Appalachian Plateau have a subtle asymmetry with the steep limb dipping toward the interior of the mountain belt. The initial Davis and Engelder (1985) assumption that the Appalachian Plateau was in a state of homogeneous Coulomb failure cannot be used to explain the asymmetry.

If Coulomb failure occurs locally and behaves according to the Davis and Engelder (1985) model, then the Appalachian Plateau asymmetry and back thrusting can be explained using the following hypothetical sequence of events. First, initial Coulomb failure and subsequent thrusting is symmetrical on both forward and backward thrust planes (Figure 9a). This thickens the section in the vicinity of the blind thrusting. Thickening acts to increase the strength of the rocks by virtue of increased normal stress across Coulomb-failure planes. The next point of failure must then be backward or forward of the thickened section

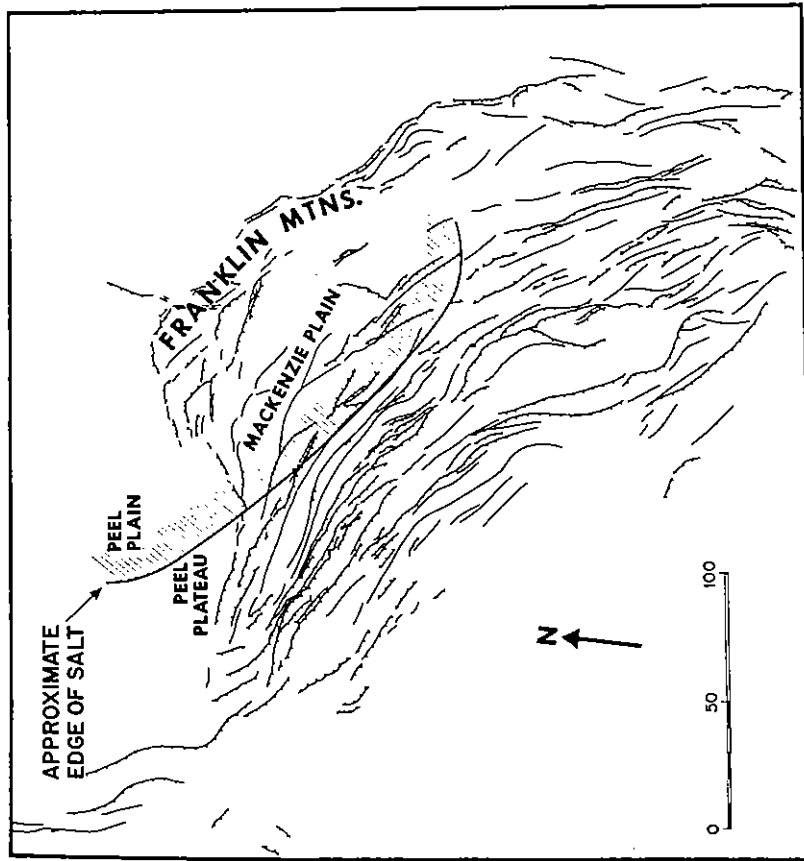


Figure 7. Map of the Franklin Mountains area (after Davis and Engelder, 1985), showing the relationship between structures and the edge of the Cambrian Saline River Formation salt.

predominantly shale section of the Plateau was affected by a penetrative strain probably reflecting a state of time-dependent viscous failure.

We noted previously that plateau structures are more symmetric than are those of the Valley and Ridge. This observation is consistent with the arguments of Section III.C.2. However, in detail many of the folds on the Appalachian Plateau are cored with blind thrust systems where the majority of the thrusts follow backward planes (Gwinn, 1964). This geometry

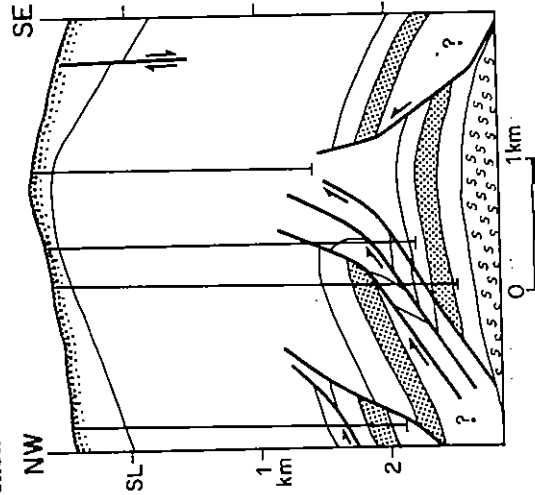


Figure 8. A cross-section through part of the Chestnut Ridge anticline in Pennsylvania (after Gwinn, 1964).

avored because of stratigraphic strength anisotropy, which should favor thrusting closer to the orientation of bedding. In addition, the section behind the frontal thrust zone has typically been thickened (and therefore strengthened) by earlier thrusting. However, if the basal detachment is extremely weak (i.e., in salt), ψ_b is very small, so the two candidate slip planes have more nearly equal dips, the wedge is much more subtly tapered, and forward-vergent thrusts should be less predominant. If folds in salt-basal fold belts are related to blind thrusts, which should face both directions, then the folds should also have relatively symmetrical form.

IV. APPALACHIAN PLATEAU DEFORMATION

A. Contrast with the Central Appalachian Valley and Ridge

Perhaps the single most distinctive aspect of the Appalachian Plateau Province is seen in large-scale map view. In the Plateau of New York and Pennsylvania, Appalachian deformation proceeds about 200 km further onto the craton than at other points along strike. This deformation is thin-skinned, and has a decollement in salt. At present, the post-salt rocks thin from 3.7 km at the southeastern end of the Plateau to 600 m to the northwest, giving a present-day wedge taper only 1.15° (Frey, 1973). It is not certain exactly how much thicker the taper was when the foldbelt was active, but it was clearly always quite narrow.

The narrow wedge taper and broad foldbelt of the Appalachian Plateau are highly suggestive of the theoretical discussion of the previous section. Several other salt-basal foldbelts share these attributes of narrow taper and unusually great breadth, including the Salt Range of Pakistan (e.g., Burbank, 1983; Jaume et al., 1985), the Franklin Mountains and the Colville Hills in front of them (e.g., Cook and Aitken, 1973), the Zagros Simply

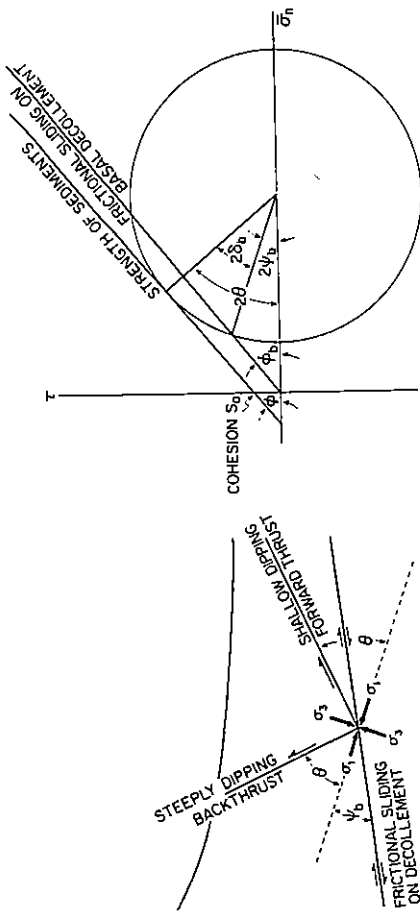


Figure 5. (a) An illustration of why, in the presence of significant decollement strength, forward vergent slip planes dip more shallowly than do the backward vergent planes. This dip difference is a direct result of the fact that ψ_b is positive. For a salt-basal fold belt, ψ_b is very small ($<1^\circ$), so thrusts and folds are relatively symmetric. (b) A Mohr-Coulomb diagram illustrating graphically the relationship between the stress orientation angle ψ and the strengths of the wedge and the basal detachment ψ_b (after Davis and Engelder, 1985).

Folded Belt (e.g., Farhoudi, 1978; Colman-Sadd, 1978), the Sierra Madre Oriental (e.g., de Cserna, 1971), and the Parry Islands Fold Belt (e.g., Price and Douglas, 1972; Fox, 1984).

The Appalachian Plateau of Pennsylvania and western New York State lies in front (to the northwest) of the Valley and Ridge Province of the Central Appalachians (Figure 6). The folds of both provinces are involved in the northward thin-skinned transport of allocthonous Paleozoic rocks. Along the boundary between the two provinces, the level of decollement steps up from Cambrian shales beneath the Valley and Ridge Province to eventually reach the Silurian Salina salt beneath the Appalachian Plateau (Figure 6).

These wavelengths are comparable in the two provinces; 8 to 16 km in the Valley and Ridge and a bit greater (8 to 32 km) in

Laboratory measurements of the deformation of salt (e.g. Carter and Hansen, 1983) indicate that for reasonable strain rates salt typically reaches its brittle-ductile transition at a depth of less than 1 km (Figure 4). Indeed, in the presence of small amounts of water, dynamic recrystallization (e.g. Spiers et al., 1986) may make salt even weaker at geologic strain rates than the low strength suggested by most laboratory measurements. Such extreme weakness seems reasonable in light of the intense shallow deformation that has been observed within salt domes and salt glaciers.

At typical depths for a basal detachment (2 to 8 km), the shear strength of salt is commonly less than 1 MPa for typical strain rates and geothermal gradients, making it between 1 and 2 orders of magnitude weaker than most other rocks (Figure 4). Anhydrite is also weak at very moderate temperatures, although not as weak as is halite. Muller et al. (1981) found that a fine-grained anhydrite from the Table Jura of Switzerland (Wandflue) is weak (<10 MPa) at low strain rates and temperatures over 100°C (Figure 4), but at a strain rate of 10^{-10} sec⁻¹ it is weak only at temperatures of over 200°C. Furthermore, they found that a less fine-grained anhydrite from Riburg required roughly 100°C higher temperatures in order to become as weak as its fine-grained counterpart at similar strain rates. The grain size appropriate for natural deformation will, of course, vary with both initial depositional character and the degree of recrystallization under strain.

It is the extreme weakness of salt that makes the structural style of overthrusting on salt so distinct from that over other types of rock (Davis and Engelder, 1985). The two mechanisms for weak-coupling decollement (overpressures and evaporites) are not mutually exclusive. Indeed, salt can form an impermeable lid for the maintenance of excess pore pressures (e.g., Fertl, 1976).

C. Effect of salt on fold belts

1. Foldbelt taper and width

A detachment in salt has a dramatic effect upon the cross-sectional taper and width in map-view of a fold-and-thrust belt. The critical taper equation (eqn. (2)) is, to a large extent, a reflection of the relative strengths of the overthrusting wedge and the stratum over which it rides. Thus, a weak detachment zone should not require a large wedge taper in order to permit overthrusting.

As we have seen in Section III.B and in Figure 4, the basal salt layer in fold-and-thrust belts which overthrust salt is typically a few km deep and is likely to be well below the depth of brittle-ductile transition in salt. For reasonable strain rates, the shear strength of the salt layer is not much greater than 1 MPa, and may be considerably less. Furthermore, for a constant strain rate, that strength is expected to be independent of the normal traction exerted by the overburden. We thus express the shear strength of the basal detachment in salt as a constant, τ_0 , whose value is likely to be less than 1 MPa. The critical Coulomb taper of such an overthrusting wedge is calculated by replacing μ_b in eqn. (2) with $\mu_b = \tau_0 / \rho g z$, giving

$$\alpha + \beta = \frac{\beta \rho g z + (1 - \lambda) \tau_0 - [2 S_0 \cot \phi (\alpha + \beta) / (\operatorname{cosec} \phi - 1)]}{\rho g z [1 + 2(1 - \lambda) / (\operatorname{cosec} \phi - 1)]} \quad (4)$$

Thus, a foldbelt needs a taper of only a few tenths of 1° in order to ride over a salt layer, if salt is as weak as eqn. (4) suggests and if the rocks of the foldbelt deform by brittle failure. Non-brittle failure in the wedge will be considered further (in Section IV.C).

Given a certain thickness of the sedimentary section, a narrower wedge taper implies that it is possible to build a broader fold belt. In other words, a reduced resistance to sliding should make it possible for the "bulldozer" to push a

Appalachians this core is presently eroded to the crystalline rocks that acted as a bulldozer pushing the clastic wedge of the Valley and Ridge and the Appalachian Plateau.

Davis et al. (1983) calculated the critical taper for the fold-and-thrust wedge of western Taiwan, making the assumption that the rock within the wedge is everywhere on the verge of shear failure according to the Coulomb failure criterion. The taper is simply the sum of the mean local slope of topography, α , and the local dip of the basal decollement, β . The magnitude of the critical taper at which the wedge can overthrust its base is governed by the balance of forces in the up-dip direction of the decollement. It is a function of the internal friction angle ϕ of the wedge, the friction coefficient at the base of the wedge μ_b , and λ , S_0 , ρ , g , and z , all of which were defined above for Eq. (1). In the limit as decollement strength becomes much less than that of the overlying sediments (which, as we shall see, is appropriate for thrusting over salt) the critical Coulomb wedge taper can be found iteratively using a relation adapted from Dahlen et al. (1984).

$$\beta \rho g z + (1-\lambda) \mu_b - \left[2S_0 \cot \phi (\alpha + \beta) / \rho g z (\operatorname{cosec} \phi - 1) \right] \\ \alpha + \beta = \frac{1 + 2(1-\lambda) / (\operatorname{cosec} \phi - 1)}{\quad} \quad (2)$$

Knowing the wedge geometry and having measured the pore-fluid pressure ratio $\lambda = 0.67$ (pore pressure $\approx 67\%$ of lithostatic) for the Taiwan fold-and-thrust belt, Davis et al. (1983) find that the frictional strength of the decollement does not have to be exceptionally low: it appears to be 80% to 85% of the strength of the overlying rocks. If the decollement in Taiwan were much weaker than 80% of the strength of the overlying rocks, then the critical taper would be smaller than its actual value of 9° . Of course, the decollement cannot be stronger than the overlying rocks, or else some other, weaker, horizon in the overlying rocks would become the layer of decollement.

III. WEAK DETACHMENT MECHANICS

A. Rock strengths

Strain in the crust takes place very slowly by human standards: a strain rate of $10^{-10} \text{ sec}^{-1}$ represents very rapid deformation, and $10^{-14} \text{ sec}^{-1}$ is more typical. In this range of strain rates and at temperatures less than about 250°C , almost all common crustal rocks deform in a brittle manner that is essentially independent of time (eqn. (1)). Byerlee (1978) pointed out that an extraordinarily wide range of crustal rocks have a very similar friction coefficient of 0.85. The only major exception that he noted was for clay minerals (some of which are often found in fault gouges), which were typically 3 or 4 times weaker. At the high temperatures of mid-crustal depths, quartz-rich rocks start to become ductile, because of thermally activated flow mechanisms. Below this brittle-ductile transition, rock strengths decrease rapidly with depth (Figure 4).

The results of experiments on dry natural rocksalt (Carter and Hansen, 1983) in the strain rate range $10^{-6} \text{ sec}^{-1} > \dot{\epsilon} > 10^{-9} \text{ sec}^{-1}$ can be fit very well by a steady-state flow equation of the form

$$\dot{\epsilon} = A (\sigma_1 - \sigma_3)^n \exp[-B/RT] \quad (3)$$

where $A = 7.6 \cdot 10^{-4} \pm 2.8 \cdot 10^{-3} \text{ sec}^{-1}$, the activation energy $B = (66.5 \pm 17.5) \cdot 10^3 \text{ kJ/mol}$, $\sigma_1 - \sigma_3$ is the stress difference, $n = 4.5 \pm 1.3$ is a constant, R is the gas constant, and T is the temperature (in $^\circ\text{K}$). Note that the strain rate is dependent upon the stress difference, $\sigma_1 - \sigma_3$, but not upon the absolute magnitude of the confining stress. In this way, ductile flow is fundamentally different than frictional strength, which is proportional to confining pressure.

its base (Figure 2). He assumed a reasonable coefficient of friction and calculated how strong the granite of the overthrust must be in order to permit a long thrust sheet to be pushed from behind. Laboratory tests suggested that granite was not strong enough for a push of this magnitude. Instead, the granite sheet would crush at the rear before slip could be initiated. Smoluchowski concluded that either all long thrust sheets must glide down hill with the help of gravity or that the assumed coefficient of friction was too large. To date, there is no evidence for widespread, downhill sliding of large thrust sheets.

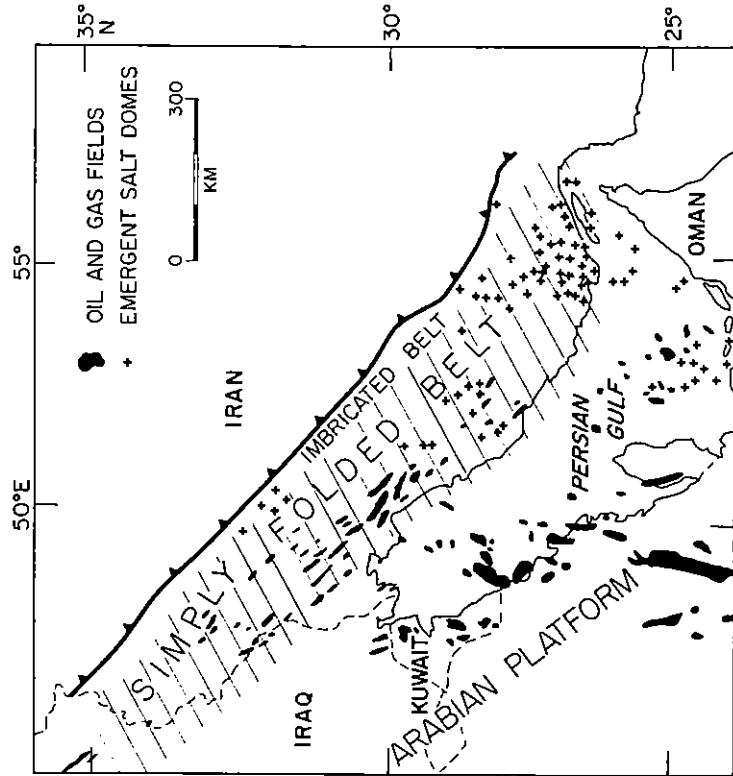


Figure 1. (a) Map (after Colman-Sadd, 1978) illustrating the relationship between salt domes, elongate oil and gas fields, and the Simply Folded Belt of the Zagros, which overthrusts salt. (b) Highly schematic cross-section through the Zagros (after Farhoudi, 1978), illustrating the complicated relationship between diapirism of salt (shaded) and folding.

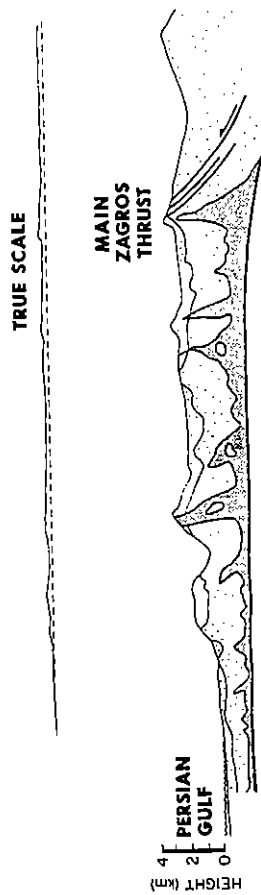


Figure 1 continued.

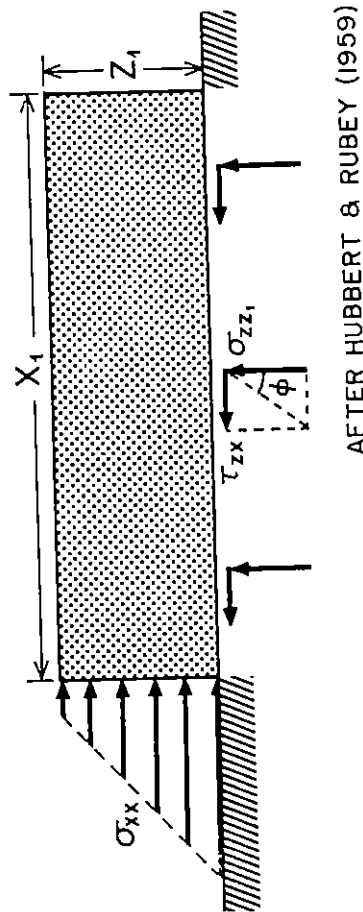


Figure 2. The balance of forces on a simple overthrust block. The frictional resistance to sliding must be balanced by a horizontal push.

This problem was partially resolved by the recognition (Hubbert and Rubey, 1959) that frictional resistance to sliding along a thrust fault is reduced in the presence of elevated pore-fluid pressures. Pore-fluid pressures support part of the normal stress but none of the shear stress, thus bringing the total stress state closer to failure. The general Coulomb criterion for the shear traction τ at failure is

$$\tau = S_0 + \mu \sigma_n (1 - \lambda) \quad (1a)$$

where $\mu = \tan \phi$ is the coefficient of internal friction, σ_n is the normal traction, S_0 is the cohesive strength, and λ , the

