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Development of the Late Mesozoic to Early Cenozoic Structures of the Eastern Great Basin²

Low-angle faults of large displacement are an important structural element of the eastern Great Basin under discussion here (Fig. 1). These faults and related structures were developed mainly during the late Mesozoic to early Cenozoic in marine and continental strata from 30,000 to 40,000 ft (9000 to 12,000 m) thick. Subsequent block-faulting (Basin and Range orogeny), erosion, sed-

imentation, and vulcanism obscured or removed these structures in many places. Enough is exposed, however, to enable us to formulate ideas of how the low-angle faults may have formed. We suggest that they were produced mainly by differential uplift and gravity sliding, and we also proffer a tentative model to explain the unique style of displacement of the low-angle faults in the western

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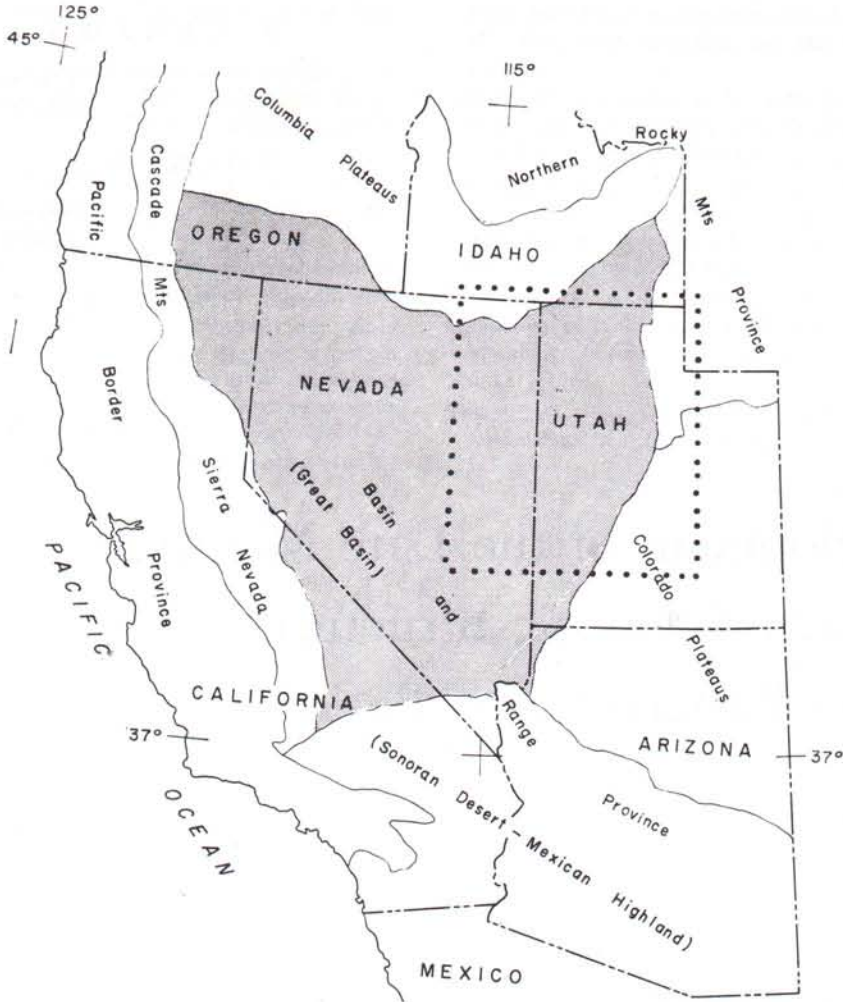


FIG. 1 Index map showing physical divisions of the western United States and outline of the subject area (dotted). Great Basin shown by gray area.

part of the area where younger or structurally higher strata are emplaced over older or structurally lower strata.

Regional Structural Geology³

The part of the eastern Great Basin that is considered here (Fig. 1) can be divided into

an eastern and western part on the basis of structural differences (Fig. 2). The two parts are separated approximately by a line that extends from the eastern edge of the Raft River Range to the southeastern edge of the Snake Range. This line is "the eastern border of known occurrences of décollement thrusting" (Misch, 1960, p. 18), and it is also the eastern boundary of Misch's "northeastern

³ The reader who is interested in the geologic history of the Great Basin before the late Mesozoic to early Cenozoic orogenesis should refer to the paper in this book by Roberts and Crittenden.

Nevada structural province." The eastern part of the area, which is unnamed, extends eastward to the eastern limit of thrusting (Fig. 2).

Low-angle faults of large displacement are the most striking pre-mid-Cenozoic structural features of this region. They were first described by Blackwelder (1910) in the Wasatch Mountains, but because they could not be traced beyond the individual mountain ranges, due to the disruption of the older terrane by the Basin and Range orogeny, their continuity along the eastern edge of the Great Basin went unrecognized for many years. Billingsley and Locke (1933) were the first to recognize the continuity of the thrust structures in the various ranges along the eastern edge of the Great Basin, and they named the whole structure the "Rocky Mountain-Great Basin thrust arc." Later it was called "the zone of Laramide thrusting" by Harris (1959) and, more recently, "Sevier orogenic belt" by Armstrong (1968).

Low-angle faults in the western part of the region were first discovered by Nolan (1935) at Gold Hill, Utah. Later, Hazzard and others (1953) described a large low-angle fault in the Snake Range, Nevada. Discovery of this fault was followed by similar finds (Misch and Easton, 1954) in the Schell Creek Range, and later Misch (1957) suggested that these faults also extended to the northern Egan Range, Cherry Creek Range, and Ruby Mountains, all in Nevada. Hazzard and Turner (1957) suggested that this fault or fault system was also present in the East Humboldt and Pequop Ranges, Nevada, the Albion Range, Idaho, and the Raft River and Grouse Creek Ranges, Utah.

Misch (1960) made the point that all of his eastern Nevada structural province was probably underlain by a décollement-type fault or fault complex, and that it was probably coextensive with and correlative with the thrust faults of the southern Wah Wah Range described by Miller (1966) and San Francisco Mountain area, Utah, described by East (1966).

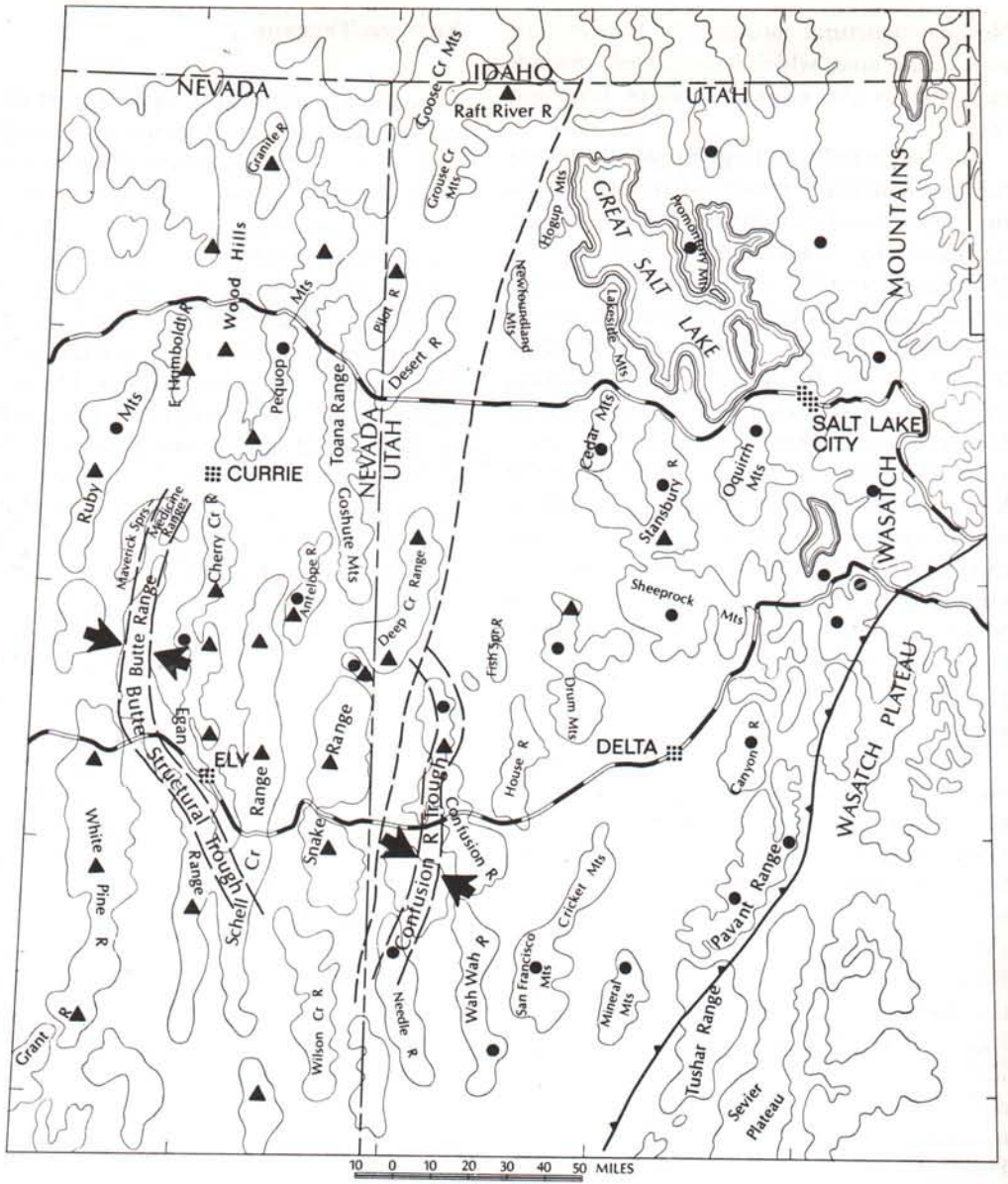
Eastern Terrane

Low-angle faults of the eastern part of the region generally have large displacement, and most of them emplace older rock over younger rock. These faults have moved a thick miogeosynclinal sequence over a thin cratonic sequence of different facies. Estimates of separation vary from as little as 12 miles (19 km) (Hintze, 1960) to as much as 75 miles (120 km) (Eardley, 1951, p. 330). Crittenden (1961) has estimated eastward movement of 40 miles (65 km) along a fault in the Wasatch Mountains. Low-angle faults with the same sense of displacement occur structurally higher within the allochthon, and most of these also emplace older rock on younger, although there are a few exceptions in the western part of the eastern area.

The allochthonous rocks of the eastern part of the area are broadly folded and may show steep dips near the frontal thrusts. Crittenden (oral communication, 1970) has mapped a large recumbent fold overturned to the east in the autochthonous rock in the Wasatch Mountains. High-angle faults within the allochthon are mainly tears associated with the higher level imbricate thrusting. In the Wasatch Mountains, rocks of presumed Paleocene age depositionally overlap a large thrust; in the Canyon Range the fault at the base of the allochthon cuts rocks that are of probable Cretaceous age—relations that seem to place the orogeny within the Cretaceous to early Paleocene.

Western Terrane

Structures of the western part differ from those to the east. Both regions contain low-angle faults of large displacement, but most of those on the west emplace younger or structurally higher rock over older or structurally lower rock. Among the few exceptions, one in the Antelope Range developed earlier than a large younger-on-older fault in



- EXPLANATION
- ▲ Younger on older low-angle faults
 - Older on younger low-angle faults
 - ↙ Eastern limit of late Mesozoic to early Cenozoic overthrusting
 - Structural contrast boundary: northeastern Nevada structural province of Misch (1960) on the west
 - Structural trough

FIG. 2 Subject area showing distribution of structural elements.

the same range. Where the major or basal décollement is exposed, as in the Snake Range, structurally higher low-angle faults are also mostly of the younger-on-older type. The allochthonous units contain normal faults, in great abundance in places, and these form series of steps, or horsts and grabens, which are truncated by the low-angle faults.

Another important feature of the western part is the general homoclinal structure of many of the mountain ranges, even those cut by low-angle faults. Young (1960, p. 166) noted that in the north-central Schell Creek Range "Folds are surprisingly scarce in the map area considering the prevalence of thrust faults," but these low-angle thrust faults are of the younger-on-older type. The only conspicuous folds are along the axial part of the Butte structural trough (Fig. 2), and these are of moderate to large wavelength and moderate amplitude.

Autochthonous rocks are present in several of the ranges in the western part of the area. The rocks exhibit a different structural style from that of the allochthon, and all are metamorphosed to some extent. In the Snake Range, Nevada, strongly rodded autochthonous quartzite and deformed marble form a gentle arch. Autochthonous Precambrian and Paleozoic strata in the Ruby Mountains are greatly attenuated, metamorphosed, and tightly folded (Howard, 1966). Similar features occur in autochthonous rocks of the Raft River Range, Grouse Creek Range, and Deep Creek Range, Utah, and in the Wood Hills, Pequop Mountains, Cherry Creek Range, and Schell Creek Range, Nevada. Metamorphism and deformation of these rocks occurred before movement on the décollement, as unmetamorphosed rock rests tectonically on metamorphic rock. We regard the folding and gentle arching of the décollement of the Snake Range reported by Hazzard and others (1953) as a late phase of the late Mesozoic to early Cenozoic orogeny.

Extent and Correlation of Low-Angle Faults

Misch (1960) inferred that the allochthonous rocks of his eastern Nevada structural province were coextensive with the allochthonous rocks of a part of the Sevier orogenic belt. It is our intent to strengthen this inference by summarizing briefly some of the geological work that has become available since Misch's (1960) paper was published. The work of Whitebread (1969) in the southern Snake Range, Nevada, of Misch and Hazzard (*in* Hose and Blake, 1970) in the central Snake Range, and of Hose and Blake (1970) in the northern Snake Range, has established essential structural and stratigraphic differences between autochthonous and allochthonous rocks in the Snake Range and immediately adjacent areas.

First, except for intrusive rocks which are present in abundance, autochthonous rocks are early Middle Cambrian or older and, second, they are generally metamorphosed and internally deformed. Additionally, high-angle faults are much less common in the autochthon than in the allochthon. On the other hand, rocks of the allochthon range from Middle Cambrian to Permian in age and are essentially unmetamorphosed but severely broken by both high- and low-angle faults. Furthermore, no intrusive rocks have been reported from upper plate rocks of the Snake Range.

In view of the profound stratigraphic and structural differences between autochthon and allochthon observed along the entire 50-mile (80 km) length and 18-mile (29 km) width of the Snake Range, we feel that the differences should extend at least a short distance beyond the range. We contend also that these differences would enable us to resolve the very important question whether the décollement of the Snake Range extends eastward beneath the Burbank Hills (a southwestern appendage to the Confusion Range, Utah) or emerges in the intervening area,

which is 5 miles (8 km) wide and almost entirely covered by younger deposits. Whitebread's (1969) detailed map shows that the single isolated outcrop of Paleozoic in the covered area, as well as the westernmost edge of the Burbank Hills, contain Devonian rock assigned to the same formation found in the easternmost part of the southern Snake Range where it is clearly part of the allochthon. Therefore it seems most probable that the décollement does indeed extend eastward beneath the Burbank Hills, for if it had emerged somewhere beneath the covered interval, the Burbank Hills would have been autochthonous and therefore most likely made up of early Middle Cambrian or older strata (Fig. 3).

Another problem of importance is whether the décollement of the Snake Range is co-

extensive with the thrust faults of the Sevier orogenic belt. Resolution of this problem is wholly dependent upon the conclusion just reached that the décollement extends beneath the Burbank Hills. Figure 3 shows that the allochthonous rocks of the Snake Range, the Burbank Hills along with the Confusion Range proper, and the Wah Wah Range form a structurally discrete and continuous unit. This unit, although cut by high-angle faults, contains only one low-angle fault that could compare in magnitude of displacement with the décollement. Miller (1966) shows that the Lower Cambrian rocks of the southern Wah Wah Range are thrust relatively eastward over strata at least as young as Middle and possibly Late Jurassic.

This thrust fault, plus two mapped in the vicinity of the San Francisco Mountains

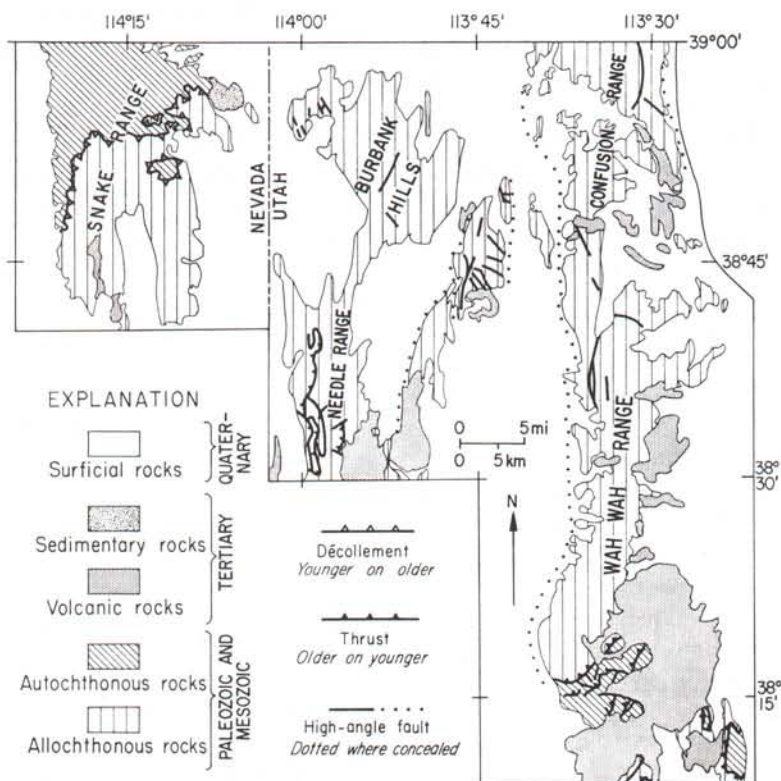


FIG. 3 Generalized geologic map of part of western Utah and eastern Nevada showing continuity of the allochthon and its relation to the autochthon.

(East, 1966), are part of the Sevier orogenic belt, and they juxtapose quite different facies of Paleozoic rocks, as is the case with thrusts to the northeast in the Wasatch Mountains. Their displacement must be considerable; furthermore, since the Wah Wah Range is in the upper plate with respect to the thrusts of the Sevier orogenic belt and, by virtue of its structural continuity with the Burbank Hills, is almost certainly upper plate with respect to the décollement, these low-angle fault surfaces must be coextensive and correlative (Misch, 1960; Hose and Daneš, 1968) (Fig. 3).

Interpretation of Contrasts in the Allochthonous Terrane

If the correlations are valid, we are faced with the problem of explaining contrasting structural styles in adjacent regions. Folding,

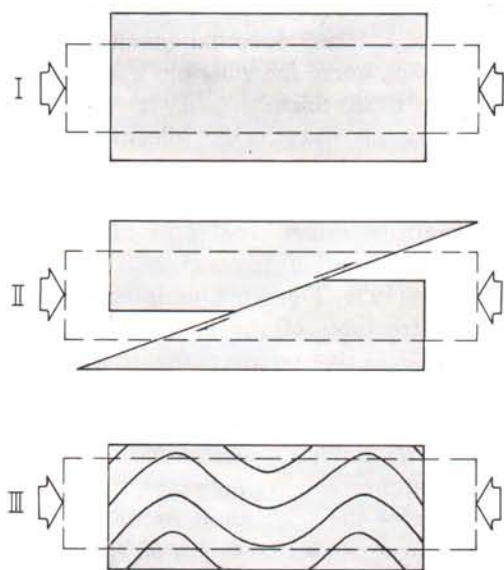


FIG. 4 Diagrams showing the ideal consequences of compressive forces applied to a slab as indicated by arrows. (I) Elastic-plastic deformation; (II) thrust-reverse fault deformation; (III) fold deformation (not to scale). (----- = original shape; — = final shape.)

thrusting, and general thickening of the section are characteristic of the eastern part of the allochthon and this could result only from horizontal compression as idealized in Fig. 4. On the west, the original section is attenuated as suggested by Fig. 5, particularly III and IV, a situation that could result only from extension. This combination of structures in one allochthonous complex is precisely what would result if the region were uparched on the west or downwarped on the east, so that an east-facing slope was formed. Under the circumstances defined by Hubbert and Rubey (1959) in which a fluid zone of greater than hydrostatic pressure is present near the base of a thick tabular mass (the developing allochthon) so that it could glide eastward, extension would prevail in the vicinity of the uparched area whereas compression would prevail at the toe. Such a combination of conditions should result in attenuation of the section on the west and thickening on the east.

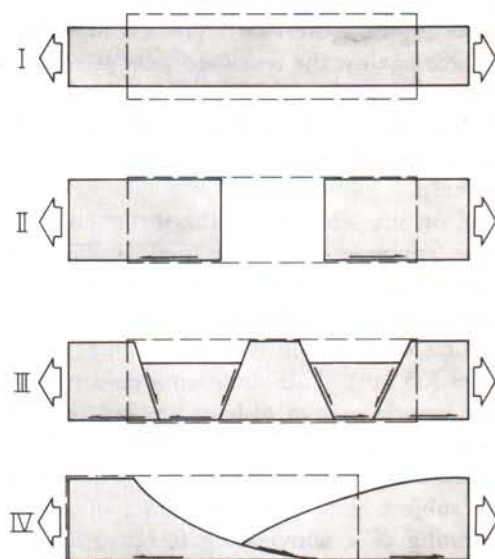


FIG. 5 Schematic diagram of effects of tensile forces applied to a slab as shown by arrows. (I) Elastic-plastic deformation; (II) pull-apart; (III) horst and graben on a basal shear surface; (IV) cycloidal shear surfaces. (----- = original shape; — = final shape.)

The combined width of the two parts of the allochthon at the present is 200 miles (320 km). If we accept 40 miles (65 km) of movement (Crittenden, 1961) on an important thrust of the frontal edge, then the original width of the developing allochthon would have been about 160 miles (260 km) or less. The thickness of the allochthon at the onset of tectonism was greater than 18,000 ft (5500 m) and perhaps locally as much as 35,000 ft (10,700 m) or even 40,000 ft (12,200 m). In view of the many lines of evidence for uniform relative eastward movement of the allochthon, a completely compressive origin would require that a push be applied to the sides as in Fig. 4 (II and III). However, application of tectonic compressive forces to this prism of strata without deformation of the autochthon is mechanically difficult to envision. One could, of course, visualize crustal (plus mantle) convergence as conventionally portrayed in sketches of tectogenes or *zwischengebirge*, and argue on this basis for compressive deformation and thickening of the rocks, which in turn would exert horizontal components of force against the rearward part of a prism of strata. Such an explanation would require severely deformed terrane in the rearward part of the allochthon—a situation that does not exist.

If, on the other hand, the thrust complex were compressive and rooted, uplift and movement along the root zone would be at least as much as the movement along the leading edge of the allochthon; that is, 40 miles (65 km). This situation would require that mantle rock or at least low crustal rock be exposed in the western part of the allochthon—also a circumstance not borne out in the subject area. The mechanics of underthrusting of a nonyielding foreland as suggested by Misch (1960, p. 41) imply that compression would have prevailed along the entire width of the allochthon. By any of these explanations, the entire allochthon should have developed internal structures that either resulted from compression or at least contained few, if any, that resulted from

extension. Certainly most of the structures of the western part of the subject area could not possibly have been formed by compression; on the contrary, the only reasonable conclusion is that they were produced by extension.

Role of Gravity

In the past, suggestions that the allochthon could have been emplaced by gravity were rejected, at least partly because no satisfactory mechanism was known that would allow gravitational movement on a small gradient. The analysis of Hubbert and Rubey (1959) circumvented that objection. But Misch (1960, p. 40) objected to a gravitational origin because he felt that tectonic denudation of the basement should have resulted. We feel, on the other hand, that total denudation would be possible only where an allochthon was so thin as to be able to support its own weight in a near-vertical wall. Such a situation is obtained in the breakaway zone of the Heart Mountain fault (Pierce, 1960, 1973) where the glide block was about 2000 ft (610 m) thick.

In a somewhat thicker allochthon, the result of extension in the western part of the allochthon would be development of horsts and grabens, rather similar to Fig. 5, III. This is the situation depicted by Rubey and Hubbert (1959, Fig. 9). Our interpretation of the structures of the allochthon in the eastern Great Basin requires that those in the western part of that area, including especially low-angle faults, resulted from horizontally applied tensile force and vertical gravitational compression. We suggest that the great thickness of the allochthon in the western part of the area (> 18,000 ft or 5500 m) must have been one of the critical factors in the evolution of the low-angle, younger-on-older faults.

A mathematical-physical explanation of the genesis of low-angle faults of the younger-on-older variety was first suggested to us by Varnes' (1962) work in the south Silverton

area, Colorado, where his theoretically derived shear trajectories compared remarkably well with the fault pattern he had mapped in the field. Varnes used a modification of Prandtl's compressed cell to derive the pattern of shear trajectories. We later found that Nye (1951) had published an analysis of shear trajectories in glaciers, very similar to the one we are presenting to explain the younger-on-older type low-angle faults of eastern Nevada and western Utah.

Our explanation is based upon the assumption that the entire allochthon was initially a large slab or prism (Fig. 6A) no more than 160 miles (260 km) across (E-W), from 18,000 to 40,000 ft (5500 to 12,200 m) thick, and at least 200 to 250 miles (320 to 400 km) long. The solution to the problem of development of low-angle faults is, however, independent of the length. We shall arbitrarily assume an initial thickness of the allochthon of 30,000 ft (9000 m).

The prism is then transformed into something similar to Fig. 6B by some indeterminate tectonic process. In a relative sense the western edge is elevated and the eastern edge depressed so that an east-facing slope is produced. Since the western edge is approx-

imately aligned with the axis of Nolan's (1943) early Mesozoic geanticline, the presence of an east-facing slope seems to be reasonably well established.

As the gradient is developed, if there exists a zone of abnormally high fluid pressure close to the base of the slab, the maximum compressive stress in the western part of the area is vertical, whereas to the east it is horizontal. In response to the vertical compression, which was induced by gravity, cycloidal shear trajectories come into being (Figs 6C and 7). The vertical compression, coupled with the tendency of the entire prism to move down gradient, causes movement along the trajectories of maximum shear, as illustrated in the left part of Fig. 6D. Faults would develop first on the west.

The mathematical analysis of the problem closely parallels that of Nye (1951), Varnes (1962), and Kanizay (1962). It is based on the following assumptions (Fig. 7):

1. The crustal block analyzed here may be approximated by an infinite slab of homogeneous, isotropic material acted upon by its own weight only.
2. Failure of the material occurs when the

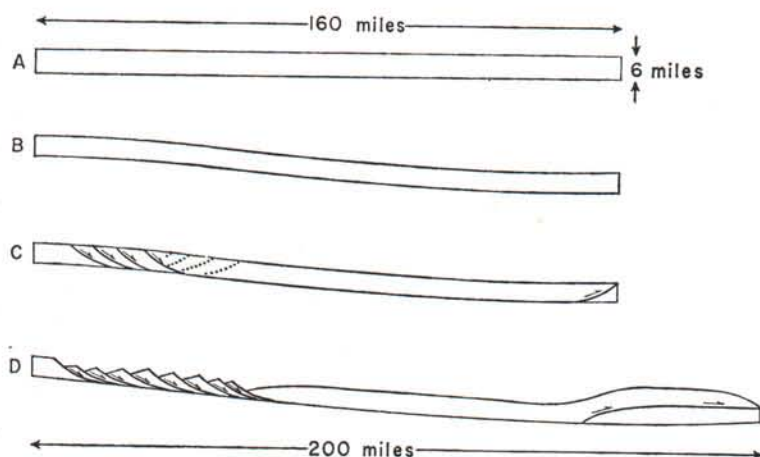


FIG. 6 Diagrammatic cross section of the eastern Great Basin showing development of younger-on-older faults. (A) Configuration of slab at outset; (B) development of gradient; (C) development of shear trajectories of cycloidal form in the rearward or active part of the allochthon; (D) final configuration showing younger-on-older faults on the west and older-on-younger faults on the east.

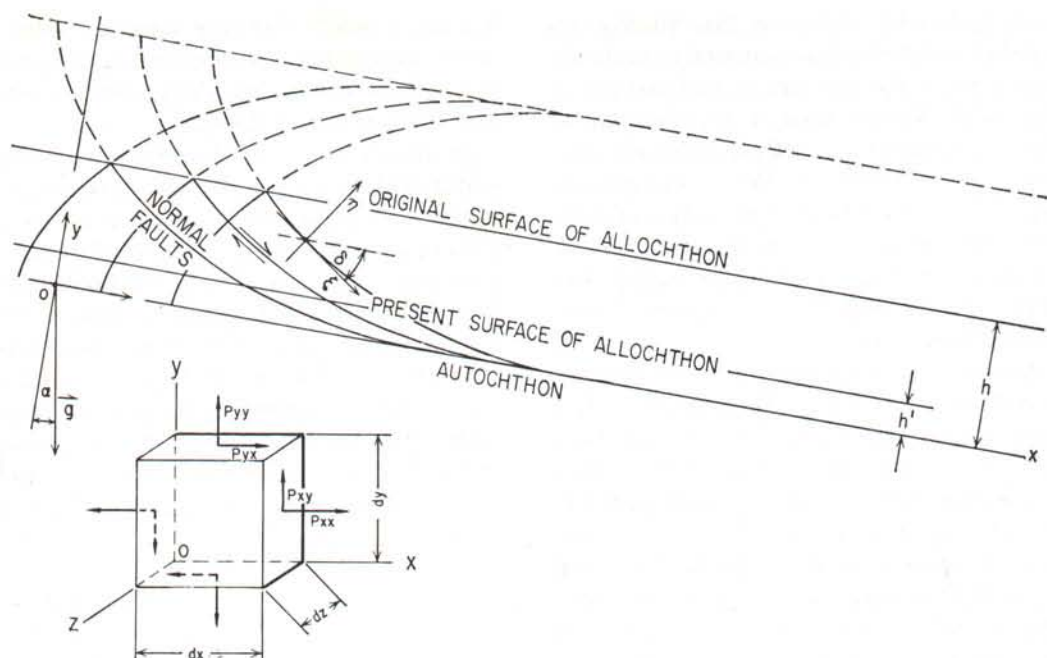


FIG. 7 Principal shear-stress trajectories within the rearward part of the postulated allochthon, and stress on a cubical element.

maximum shear stress reaches a constant value k characteristic of the material (Von Mises's criterion).

3. Failure follows a trajectory of maximum shearing stress.
4. The problem is independent of the coordinate Z .

The only body force acting on a particle is its weight; therefore

$$\frac{\partial P_{xx}}{\partial x} + \frac{\partial P_{yx}}{\partial y} = -\rho g \sin \alpha \quad (1)$$

$$\frac{\partial P_{xy}}{\partial x} + \frac{\partial P_{yy}}{\partial y} = \rho g \cos \alpha \quad (2)$$

where ρ is the density of the material, g the gravity acceleration, x the cartesian coordinate along the slab, y at right angles to the slab (positive sense up), and α the dip of the slab.

Solving Eqs. 1 and 2, setting the stress components equal to zero at the free surface,

$y = h$; and setting $P_{xy} = k$ at the bottom of the slab, $y = 0$ yields

$$P_{xx} = a(y - h) + f(y) - f(h) \quad (3)$$

$$P_{xy} = -\rho g \sin \alpha (y - h) = k \left(1 - \frac{y}{h}\right) \quad (4)$$

$$P_{yy} = \rho g \cos \alpha (y - h) = \frac{k}{\tan \alpha} \left(\frac{y}{h} - 1\right) \quad (5)$$

where a is an arbitrary constant and f is an arbitrary function.

Applying assumption 2 and the rules of tensor transformation, we are now able to obtain via a series of equations (5a to 5j) given in the Appendix the following mathematical expressions for the shape of the fault surfaces:

$$\frac{dy}{dx} = \frac{2 - y/h}{\sqrt{(y/h)(2 - y/h)}} \quad (6)$$

or

$$\frac{dy}{dx} = \frac{-y/h}{\sqrt{(y/h)(2 - y/h)}} \quad (7)$$

Both Eq. 6 and Eq. 7 represent families of cycloids. One family has the cusps at $y = 0$, pointing downward; the other, at $y = 2h$, pointing upward. Every trajectory of one of the families intersects every trajectory of the other family at right angles, and both intersect the free surface, $y = h$, at an angle of 45° .

The foregoing analysis shows how cycloidal shear trajectories could indeed be generated and that, in view of the assumed plastic nature of the prism and the east-sloping gradient, movement by body forces should take place. In the western part of the allochthon this movement would be of the type that would move structurally higher (younger) rock over structurally lower (older) rock. For the case of the eastern Great Basin we conclude that most of the upper portion of the allochthon has already been eroded since the average observed dip of the fault surfaces is gentle to flat.

Movement is possible only on a slope, and the minimum gradient, according to the work of Hubbert and Rubey (1959), is a function of the ratio λ of fluid pressure at the base of the allochthon to the weight of rock above. As the fluid pressure approaches the geostatic pressure ($\lambda \rightarrow 1$), the required gradient becomes less. Raleigh and Griggs (1963) have pointed out that it is necessary to consider the fact that energy has to be expended to force the leading edge of an allochthon up a ramp or rise, a problem not considered by Hubbert and Rubey (1959). This requires a slight increase in the gradient, or an increase in λ , or a slight push from the rear.

In the westernmost part of the developing allochthon the segment of rock resting on a hemicycloidal shear trajectory is resting on an average slope of about 22° . If λ along this shear surface is reasonably high, the block of rock resting on the cycloidal surface would exert a considerable eastward force on the back end of the allochthon, probably not enough to fully account for the extra force needed to push the leading edge up the ramp but certainly enough to minimize the increase

in gradient called for by Raleigh and Griggs.

As the allochthon moves downslope it creates a mass deficiency in the hinterland and mass surplus along the leading edge as it overrides the autochthon. The net effect in the rear would be uplift of the autochthon to provide a slight local gradient increase. Along the leading edge the increased mass would cause isostatic adjustment that would have the effect of reducing the angle of the slope against which the front part of the allochthon would move. The combined effects would make movement of the allochthon easier.

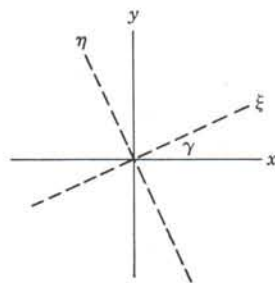
Appendix

According to the rules of transformation of components of a tensor,

$$\begin{aligned} P_{\xi\eta} \equiv & \cos(\xi, x) P_{xx} \cos(x, \eta) \\ & + \cos(\xi, x) P_{xy} \cos(y, \eta) \\ & + \cos(\xi, y) P_{yx} \cos(x, \eta) \\ & + \cos(\xi, y) P_{yy} \cos(y, \eta) \quad (5a) \end{aligned}$$

where ξ, η are an arbitrary (local) orthogonal coordinate system. Let the angle between x and ξ be γ ; then

$$\begin{aligned} P_{\xi\eta} = & -P_{xx} \cos \gamma \sin \gamma \\ & + P_{xy} (\cos^2 \gamma - \sin^2 \gamma) \\ & + P_{yy} \sin \gamma \cos \gamma \\ = & -\frac{P_{xx} - P_{yy}}{2} \sin 2\gamma + P_{xy} \cos 2\gamma \quad (5b) \end{aligned}$$



If $P_{\xi\eta}$ along such a direction is to reach a maximum value, then it must vanish along a

plane inclined 45° to the former:

$$\begin{aligned} 0 &= -\frac{P_{xx} - P_{yy}}{2} \sin 2(\gamma + 45^\circ) \\ &\quad + P_{xy} \cos 2(\gamma + 45^\circ) \\ &= \frac{P_{xx} - P_{yy}}{2} \cos 2\gamma + P_{xy} \sin 2\gamma \end{aligned} \quad (5c)$$

so that

$$\tan 2\gamma = -\frac{P_{xx} - P_{yy}}{2P_{xy}} \quad (5d)$$

Therefore

$$\sin 2\gamma = -\frac{P_{xx} - P_{yy}}{\sqrt{4P_{xy}^2 + (P_{xx} - P_{yy})^2}} \quad (5e)$$

and

$$\cos 2\gamma = \frac{2P_{xy}}{\sqrt{4P_{xy}^2 + (P_{xx} - P_{yy})^2}} \quad (5f)$$

We substitute (5e) and (5f) in (5b) and use assumption 2:

$$\begin{aligned} P\xi\eta &= \frac{1}{2} \frac{(P_{xx} - P_{yy})^2}{\sqrt{4P_{xy}^2 + (P_{xx} - P_{yy})^2}} \\ &\quad + \frac{2P_{xy}^2}{\sqrt{4P_{xy}^2 + (P_{xx} - P_{yy})^2}} \\ &= \frac{1}{2} \frac{4P_{xy}^2 + (P_{xx} - P_{yy})^2}{\sqrt{4P_{xy}^2 + (P_{xx} - P_{yy})^2}} = k \end{aligned} \quad (5g)$$

or

$$(P_{xx} - P_{yy})^2 + 4P_{xy}^2 = 4k^2 \quad (5h)$$

Next, we transform (5d):

$$\begin{aligned} \tan 2\gamma &= \frac{2 \tan \gamma}{1 - \tan^2 \gamma} \\ &= -\frac{P_{xx} - P_{yy}}{\sqrt{4k^2 - (P_{xx} - P_{yy})^2}} \end{aligned} \quad (5i)$$

or

$$\begin{aligned} \tan \gamma &= \frac{\sqrt{4k^2 - (P_{xx} - P_{yy})^2} \pm 2k}{P_{xx} - P_{yy}} \\ &= \frac{\partial y}{\partial x} \end{aligned} \quad (5j)$$

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