The Motion of Thrust Sheets

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The regional average basal shear stress τ of a thrust sheet of thickness H is equal to the down-surface slope stress $\rho gH\alpha$. Thrusts always move in the direction of surface slope α, even if they are moving up the dip β of the base. The sole thrust beneath the Canadian Rockies had τ of the order of 5×10^6 Pa (50 bars). Listric normal faults in the main ranges coincided with a reversed sense of τ. A dimensionless number gives the relative magnitude of compressive surface to gravitational forces in the formation of a thrust. The general strength of the rock imposes severe restrictions on the magnitude of surface forces, and gravitational forces dominate in the emplacement of entire thrust sheets, although compressive surface forces are important in the toes. If gravitational forces are dominant, continent-continent collision is not required to produce an orogeny, but significant surface slopes are. An important cause of such slopes is rapidly uplifted magmatic arcs. Piggyback stacks of imbricates can be interpreted in terms of the evolution of paleoslope. Ophiolite thrust sheets have affinities with thrusts made of shelf and slope sediments and may be treated geometrically and mechanically in similar ways.

INTRODUCTION

After their discovery in the late 1800's it was quickly demonstrated that thrust faults move very long distances, some more than 100 km. The early workers were impressed with the huge volume of rock involved in thrust sheets and with the fact that the individual sheets remained remarkably coherent and undeformed despite the long distances that they travelled. They learned that thrust faulting occurs throughout mountain chains and that it is one of the most important orogenic processes.

Sometime before the second world war the mechanism of moving thrusts by 'gravity gliding' became accepted. This theory has several variations, but at its simplest it states that (1) thrusts move entirely under the influence of gravity, (2) at the time of motion the thrust planes dipped downhill in the direction of motion, and (3) that at their trailing edge the thrusts cut upsection towards the surface. This gravity gliding theory was widely adopted, particularly amongst Alpine geologists actively working with thrusts.

An important alternative to this gravity gliding theory seems to have first been proposed by *Bucher* [1956, 1962]. In a series of roughly scaled experiments he noticed that model mountain chains 'spread out' under gravity from regions of higher topographic elevations, producing remarkably natural-looking thrusts and recumbent folds.

Some of the most illuminating information on thrust faults in recent years has come from field work, particularly in the Canadian Rockies foreland thrust belt. In a review and evaluation of some of this field evidence, *Price and Mountjoy* [1970] found it virtually impossible to have the thrust planes dipping downhill or to cut upsection at their trailing edges, and they rejected the theory of gravity gliding [see also *Price*, 1971]. They resurrected the theory of 'gravity spreading,' which states in brief that the mass of rock making up the thrust belt will under the influence of gravity spread sideways towards the craton, driven by the difference in surface elevation between the interior of the mountain belt and the craton. This theory implies (as does the older gravity gliding) that horizontally directed surface forces are insignificant on a regional scale.

There may well be some thrust sheets which have moved into position by gravity gliding, although I have not been convinced by field examples published to date. But this problem does not affect the main question of this paper: Is there some way of relating the surface slope and thickness of the thrust sheet to the magnitude of the shearing stress at the base?

Is it possible to prove the relative importance of surface versus gravitational forces? We shall see that the overriding factor in determining the gravitational forces acting on a thrust belt is the regional surface slope. If this is so, is it possible to use a stack of thrust sheets to reconstruct the paleogeography at the onset of deformation? Some of the largest thrust sheets place oceanic crust and mantle onto continental crust—the ophiolite complexes. Is the mechanics of the emplacement of ophiolites different from any other thrust faulting?

THRUST FAULTS AND MATERIAL PROPERTIES

Concepts of friction along the sliding thrust play an essential role in most theories. Conventional sliding friction has long been known to make thrusting impossibly difficult [*Smoluchowski*, 1909]. *Hubert and Rubey* [1959] proposed that the resistance caused by sliding friction could be reduced by a pore pressure, and this idea has been widely adopted [e.g., *Hsu*, 1969*a*, *b*; *Forristall*, 1972]. An alternative suggestion is to have viscous sliding along the thrust surface [*Smoluchowski*, 1909; *Kehle*, 1970]. Thrust sheets were treated as perfectly plastic bodies in an important but frequently overlooked treatment by *Goguel* [1948].

But there is little published observational evidence on the physical processes which were operating along a sliding thrust fault. How is one to choose between viscous sliding on the one hand and frictional sliding with some variable amount of pore pressure on the other, or, for that matter, some other totally different sliding law?

Rocks undergoing natural deformation demonstrate either linear or nonlinear flow laws depending upon deviatoric stress, temperature, grain size, and various chemical effects [e.g., *Elliott*, 1973*b*]. There is reason to believe that sliding surfaces also show this kind of behavior, and these sliding laws are the subject of separate papers [*Elliott*, 1973*a*, 1975]. It seems essential to separate conclusions based largely upon the statics of thrust sheets from results which depend upon assumptions about special laws of sliding along the base of thrust sheets.

In this paper we shall try and use as few assumptions as possible about the material properties. We shall see that considerable progress can be made by knowing only the shear 'strength' under natural conditions of the rocks making up the thrust sheet. This strength could be either a ductile yield or a fracture; the mode of deformation is not important. What we require is the lowest possible upper bound to the shearing stress τ_m which the rocks must sustain for the length of time necessary to emplace a thrust sheet.

1. Many experiments on specimens at fairly high strain rates ($\dot{e} \gtrsim 10^{-7}$ s^{-1}) suggest laboratory fracture strengths for incompetent evaporites, shales, and coals of $\lesssim 10^7$ Pa (100 bars, e.g., *Heard and Rubey* [1966]). These strengths must be extrapolated downward because of much lower geological strain rates ($\dot{e} \sim 10^{-14}$ s^{-1}). Laboratory fracture strengths of massive greywacke have been extrapolated to geological conditions of strain rate, temperature, pore pressure, and mean normal stress by *Brace et al.* [1970]. They found for the greywacke $\tau_m < 10^8$ Pa (1 kbar), but they noted that even minor quantities (~10%) of weak rock could reduce the strength of such a mixed greywacke-shale terrain nearly to the value of shale. This in essence is the result of scaling from the centimeter laboratory specimen to the geological dimensions of tens of kilometers.

2. A mechanism exists which enables a low value of regional shearing stress to be magnified locally to values sufficiently high to cause fracture. These are the local stress concentrations at the tip of a growing thrust, which may reach an order of magnitude greater than any regionally imposed shear stress [*Elliott*, 1975].

3. Another estimate of τ_m may be made with the geodetic and heat flow data from the transform faults of the San Andreas system in California. The product of the average shear stress along a fault and mean rate of slip is the rate of mechanical work. California geodetic data demonstrate that slip along the transform faults occurs mainly by creep with a comparatively small amount of slip producing earthquakes [*Savage and Burford*, 1973]. Hence most of the mechanical work is converted into heat, with little transport of energy away from the fault as elastic waves. The heat produced during faulting can be used to estimate the average shearing stress. Since several million years are needed in order to see the effects of changes in heat flow, the deformation stresses calculated from heat flow data will be long-term average stresses, exactly the sort required.

The heat flow anomalies over several active faults in California are very smooth, indicating that in the upper 20 km of the faults there is low heat generation [*Brune*, 1974]. These heat flow anomalies imply a time-averaged stress $\tau_m \lesssim 2 \times 10^7$ Pa (200 bars).

This information is all from one portion of one transform fault province. Is it possible that the mechanical processes operating along conservative plate margins are quite different in kind or degree from those operating in foreland thrust belts? Obviously, each tectonic province should be studied as a separate entity, but to the best of my knowledge there is no comparable geodetic, fault creep, and heat flow data for any active foreland thrust belt.

Seismic maps show that currently active foreland thrust belts, such as the eastern Andes, have appreciably less seismic activity than other tectonic provinces such as spreading ridges, transform faults, or subduction zones. Could this be a consequence of a lower time-averaged stress along thrust faults than on transform faults, so that an even higher percentage of thrust fault motion is via creep?

If so, one should expect even smaller heat flow anomalies over thrusts than over transform faults. *Hsu* [1969*a*] states, without citing evidence, that no significant heat flow anomaly is produced over thrust faults. If the stresses operating during thrusting are sufficiently small that no detectable heat flow anomaly is produced, then the long-term average stresses are capable of satisfying such a condition $\tau_m \lesssim 10^7$ Pa (100 bars) [*Brune et al.*, 1969].

For these three reasons I feel that a long-term upper limit to the strength of the rock in a thrust sheet is about 2×10^7 Pa (200 bars).

BASAL SHEAR STRESS AND THE LONGITUDINAL STRESS EQUATION

As in many cases in geology, velocities are so small that inertia is negligible in comparison to the magnitude of other terms, and the general equations of motion may be written as the equations of creeping motion or static equilibrium,

$$\frac{\partial \sigma_{11}}{\partial x_1} + \frac{\partial \sigma_{12}}{\partial x_2} - \rho g \sin \alpha = 0 \qquad (1a)$$

$$\frac{\partial \sigma_{22}}{\partial x_2} + \frac{\partial \sigma_{21}}{\partial x_1} + \rho g \cos \alpha = 0 \qquad (1b)$$

Here σ_{ij} are the components of the stress tensor, the first and second suffixes being the directions of traction and outward normal to a surface, respectively. Tensile stresses and strains are positive. Density ρ is assumed to be constant, and g is the gravitational acceleration. The x_1 coordinate is parallel to the surface, dipping at an angle α from the horizontal (Figure 1).

Assumed boundary conditions, natural to use in a geological context, are that the top surface is stress free and that the flow is essentially two dimensional or in a state of plane strain.

Equations (1*a*) and (1*b*) and boundary conditions apply to any material undergoing slow creeping motion or in a state of static equilibrium.

In this paper we will be neglecting the longitudinal stress gradients $\partial\sigma_{11}/\partial x_1$ and $\partial\sigma_{21}/\partial x_1$, but how can we justify this? This important question takes up a substantial portion of this paper, and we shall see that there are certain geological conditions which must be satisfied. Three main approaches will be used to demonstrate the approximation:

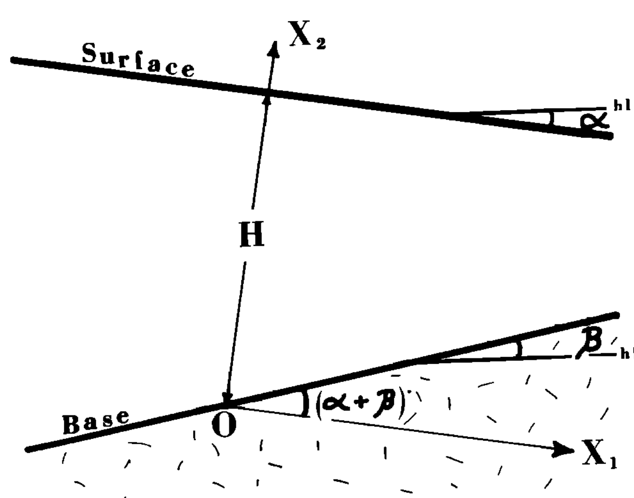

Fig. 1. Portion of a thrust sheet with surface and basal slopes α, β measured from the horizontal. Rectangular Cartesian coordinates have origin O resting on the base of the thrust; x_1 parallels the surface and lies at an angle $(\alpha + \beta)$ to the base. The thickness H of the sheet is measured along the x_2 coordinate.

1. Thrust faulting will be seen as a consequence of having the mean normal stress approximately equal to the weight of overburden and the base of the thrust sheet undergoing a large component of simple shear.

2. Thrust faulting will be examined through an exact 'longitudinal stress equation' which includes the longitudinal stress gradients. Possible geological effects of these stress gradients will be commented on, and conditions will be formulated when the longitudinal stress equation approximates our simpler equation.

3. Thrust faulting is frequently attributed to a stress state arising from a horizontal compressive push. A simple and explicit stress state will be examined as an example of combined compressive surface and gravitational forces. This stress state includes as a special case the system proposed by *Hubbert* [1951] and *Hafner* [1951]. Later in the paper we shall describe the conditions under which these compressive surface forces are negligible in comparison to the gravitationally induced forces.

Let us start by writing (1*a*) in terms of mean $\sigma_m = (\sigma_{11} + \sigma_{22} + \sigma_{33})/3$ and deviatoric σ_{11}' stresses. After integrating from the stress-free surface down to a depth H

$$\int_0^H \frac{\partial \sigma_m}{\partial x_1}\, dx_2 + \frac{\partial \sigma_{11}'}{\partial x_1}\, dx_2 + \sigma_{12} - \rho g H \sin\alpha = 0 \qquad (2)$$

Allow the mean normal stress σ_m to approximate the weight of overburden. This assumption of the 'lithostatic' state is of course almost universal throughout metamorphic petrology, $\sigma_m \simeq -\rho g H$.

The validity of this approximation depends upon the ability of the rock to withstand shear stress. By a depth of 760 m the rock strength equals the lithostatic value ($\tau_m \sim 2 \times 10^7$ Pa $\sim \rho g H$), and typical thrust sheets have $H \sim 10$ km, where the rock strength is but a fraction of the weight of overburden ($\tau_m < 10\%\rho g H$). We now restrict ourselves to thrust faults which are at sufficient depths H that

$$\frac{\tau_m}{\rho g H} < 1$$

so that $\sigma_m \simeq -\rho g H$. By substituting into the first term in (2),

$$\int_0^H \left.\frac{\partial \sigma_m}{\partial x_1}\right|_{x_2} dx_2 \simeq 0$$

Consider the conditions under which the second term in (2) is negligible with respect to the third. This will be so whenever

$$\frac{\sigma d_{11}'}{\sigma_{12}} \frac{H}{X_1^c} \ll 1$$

Here σ_{11}' is a characteristic value of the longitudinal deviatoric stress averaged over a length X_1^c. Now from the Mohr diagram (Figure 2),

$$\frac{\sigma_{12}}{\sigma_{11}'} = \tan 2\theta^c$$

A convenient averaging length $X_1^c \gtrsim 5H$, and in this case the second term in (2), is less than 10% the value of the third term whenever the angle θ between x_1 and the maximum principal stress is $\theta \geq 32°$.

We now are left only with the last two terms in (2)

$$\sigma_{12} \simeq \rho g H \sin\alpha \qquad (3a)$$

and similarly from (1*b*)

$$\sigma_{22} \simeq -\rho g H \cos\alpha \qquad (3b)$$

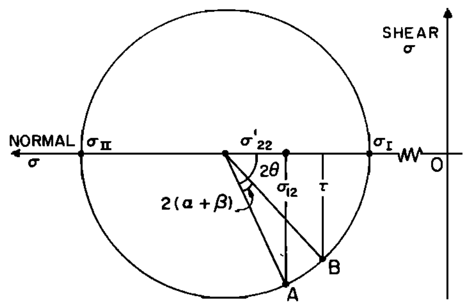

Fig. 2. Mohr diagram of stress state at point O on Figure 1. Principal stresses are σ_I, σ_{II}. Point A represents stresses σ_{12}, σ_{22}' on a plane through O parallel to x_1. In plane strain, $|\sigma_{22}'| = |\sigma_{11}'|$. Point B represents the stresses, such as τ, on a thrust plane passing through O. If the base of the thrust sheet were exactly undergoing a simple shear, $\theta = 45°$.

The neglect of all longitudinal stress gradients implies that thrust sheets in some senses approximate ductile deformation zones. Indeed, the necessary condition $\theta \geq 32°$ requires that the base of thrust sheets undergo a substantial component of simple shear strain, and this may be tested by field observations.

This basal shear stress τ_b along the thrust sheet may now be determined from the Mohr diagram (Figure 2) as

$$\tau_b = \sigma_{12} \frac{\sin 2(\theta - \alpha - \beta)}{\sin 2\theta}$$

Since $(\alpha + \beta) \sim 5°$ and $\theta \geq 32$, then from (3*a*),

$$\tau \simeq \rho g H \alpha \qquad (4)$$

We will refer to the product $\rho g H \alpha$ as the 'down-surface slope stress' and τ as the 'basal shear stress.'

A very important point emerges from this analysis. It is the surface slope which determines the magnitude and sense of the shearing stress at the base, and not the dip of the base. In fact, we see that the surface and base do not even have to dip in the same sense for the simple equation (4) to hold as a good approximation. The mass moves in the direction of surface slope, even if this means 'up' the basal slope. This is the governing equation behind the gravity spreading model.

But is there not some maximum basal slope above which no reasonable value of surface slope can drive the thrust forward? If the thrust sheet is to move forward entirely by gravitational forces, it must lose potential energy as it advances. We shall examine these restrictions later in this paper.

It may be helpful to visualize thrusting as a problem in the hydraulics of open channel flow. Equation (4) is a relation between the basal shear stress τ, the 'elevation head' H, and the 'hydraulic gradient' α. The thrusts always flow down the hydraulic gradient α.

The simple relation (4) has been applied in glaciology for 25 yr. *Orowan* [1949] appears to have been the first person to derive an equation of this form, but he assumed the special case of equal surface (α) and basal (β) slopes. *Nye* [1952] presented a somewhat different method of approximation to arrive at the same result as in this paper. No very special material properties were assumed by either *Orowan* [1949] or *Nye* [1952]. Even though glaciers are ductile solids deforming very near their melting point, whereas large portions of thrust sheets are deforming at less than half their melting point, the equations describing the statics of glaciers and thrust sheets are identical. The key restriction is that the depth be such that the shear strength τ_m of the rock (or ice) is much less than the mean normal stress.

Let us now return to the questions surrounding our neglect of the longitudinal stress gradients. We pointed out that an exact solution exists called the longitudinal stress equation which includes these stress gradients $\partial\sigma_{11}/\partial x_1$ and $\partial\sigma_{12}/\partial x_1$. Thorough and extended treatments are readily accessible in the literature because in the past few years there have been major attempts by four glaciologists [*Robin*, 1967; *Collins*, 1968; *Nye*, 1969; *Budd*, 1970*a*, *b*, 1971] to deduce the longitudinal stress equation entirely from first principles with as few approximations as possible. The most powerful and general solution is that of *Budd* [1970*a*], who derived his result by integration of the equations of creeping motion (equation (1)) starting from the stress-free surface down through the glacier to its base. He derived an exact and general integral equation without any assumptions about the material properties. This exact longitudinal stress equation [*Budd*, 1970*a*, equation (17)] involved 14 terms and can be simplified for small average surface slopes ($\alpha \lesssim 14°$) when the maximum deviation in height h from a mean straight line surface is sufficiently small ($h/H \ll 1$) to get a longitudinal stress equation

$$\tau + 2G + T = \rho g H \alpha \qquad (5)$$

This equation may be regarded as a simple force balance. Consider a passive 'dye-marked' columnar element drawn from the surface through the thrust sheet of unit height and unit area on the base (Figure 3). The forces acting in the longitudinal direction on the unit areas of the column are the down-surface slope stress $\rho g H \alpha$ due to the weight of the column, the basal shear stress τ, the force acting across the column $2G$ which overcomes the resistance of the mass to extension or compression, and the force T which must overcome the resistance to internal slip (Figure 3). Since the column is in creeping motion or static equilibrium, we can balance all these forces to arrive at (5).

The physical importance of the term $(2G + T)$ emerges from an interpretation due to *Orowan* [1949]. If we integrate all the forces acting along a thrust sheet from some point ($x_1 = 0$) to the toe (x_1), then

$$F = \int_0^{x_1} \rho g H \sin\alpha \, dx_1 - \tau x_1 = \int_0^{x_1} (2G + T) \, dx_1$$

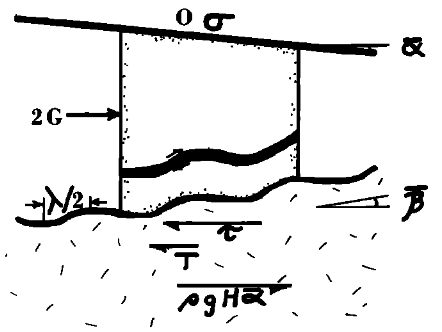

Fig. 3. Irregular thrust sheet has average surface dip $\bar{\alpha}$ and average dip of the base β. Irregularities in the base with half-wavelength $\lambda/2$ set up parallel folding within the interior of the sheet. The column with stippled boundaries has average (unit) height equal to the average width. The surface is stress free, and only the forces acting in a longitudinal direction are shown acting on the unit column. These forces are a resistance T to being flexed or folded, a resistance to extension and contraction G, average values of basal shear stress $\bar{\tau}$, and down-surface slope stress $\rho g H \bar{\alpha}$. These forces must all balance (equation (5)).

A longitudinal force F exists in general which varies from point to point along the thrust. Wherever some part of the thrust has a thickness $H > \tau/\rho g \alpha$, then this portion of the thrust exerts either a compressive or tensile force upon some place ahead or behind whose thickness $H < \tau/\rho g \alpha$. The effect of this longitudinal force is to maintain the various parts of the thrust sheet in continuous slow motion.

In a more exact sense the 'variational stress' T and the 'stress deviator gradient' $2G$ are defined by *Budd* [1970*a*] as

$$T = \iint \frac{\partial^2 \sigma_{12}'}{\partial x_1^{\,2}} \, dx_2 \, dx_2 \qquad G = \int \frac{\partial \sigma_{11}'}{\partial x_1} \, dx_2$$

Here the integration is from the surface down to the base.

Obviously, T and $2G$ reflect the longitudinal stress gradients, but it is not obvious what the physical implications of these terms are. It is worth pointing out in a simplified, intuitive fashion that several structural features within thrust sheets can be accounted for in terms of T and $2G$. Imagine a hypothetical situation with the base of the thrust sheet perfectly lubricated so that the basal shear stress term does not exist, $\tau = 0$. The moving mass must mold itself into the bumps and obstacles at the base, and an internal resistance to contraction, extension, bending, and flexing of the sheet must be overcome. This internal deformation sets up a resistance to forward motion of the thrust, and T and $2G$ are the categories of these resisting stresses.

We saw that $(T + 2G)$ reflects changes in overall thickness of the sheet, and now we will see that T and $2G$ separately may be related to undulations in the base and in the surface.

The column undergoes flexural slip as it moves over bumps in the base producing a shear stress resistance T which must be overcome. Over sinusoidal obstacles in the base the shear slip reaches a maximum at inflection points. These inflection points are spaced every half wavelength $\lambda/2$. A short distance between inflection points produces a large amount of slip, so for a constant amplitude, $T \propto 2/\lambda$. Consequently, if the wavelength is long enough, then T is negligible, but for short wavelengths, T may be a dominant term. The geological implication of a high value of T is many short wavelengths parallel or kink folds.

The net normal stress difference acting on the sides of the column is $2G$. If $2G$ is negative (compressive) over a sufficient distance along the thrust, then longitudinal shortening of the thrust sheet will occur by contraction faults, stylolite surfaces which are normal to the movement direction, and, again, parallel or kink folds. Similarly, a positive $2G$ causes lengthening of the thrust sheets by means of extension faults. Contraction and extension faults are widespread within thrust sheets [*Norris*, 1958]. The net normal stress difference $2G$ may be visualized as the normal stress necessary to extend or shorten the sheet as it moves over undulations and will exist even if the basal slip and interbed slip were both perfectly lubricated (i.e., $\tau = 0 = T$).

As a thrust moves over a sinusoidal bump, $2G$ will be negative over the synclines and positive over the anticlines, so that periods of compression will alternate with periods of extension, and the net contribution will cancel out.

Another interpretation of T and $2G$ is possible in terms of local variations in surface slope.

Instead of actual values at a particular spot of the surface slope α and basal shear stress τ define regional average values $\bar{\alpha}$, $\bar{\tau}$, and local deviations from the average values α', τ', where

$$\alpha = \bar{\alpha} + \alpha' \qquad \tau = \bar{\tau} + \tau'$$

Equation (5) may now be rewritten as

$$\rho g H \bar{\alpha} + \rho g H \alpha' = 2G + \bar{\tau} + \tau' + T$$

All four terms G, α', τ', and T reverse their signs and vary in a periodic fashion over bumps. Thus the net value of all of these terms is zero when they are integrated over a longitudinal distance x_1 which is greater than several wavelengths. Thus the regional value of the basal shear stress $\bar{\tau}$ over a distance $x_1 > 5\lambda$ is $\bar{\tau} \simeq \rho g H \bar{\alpha}$. The simple equation (4) is recovered and is clearly seen to involve only long-range average values.

Assuming that the variation in surface is sufficiently smooth, *Budd* [1971] showed that

$$\begin{aligned} G &= \rho g H \alpha' + T \\ G &\simeq \tau' \end{aligned} \tag{6}$$

Equation (6) provides a way of measuring local deviations from regional values. Notice that (4) and (6) are a set of simultaneous partial differential equations which, by taking their sum, can be seen to be equivalent to (5).

If the obstacles and bumps have sufficiently long wavelength $[(\lambda/2) \geq 2H]$, then T is negligible and $G \simeq \rho g H \alpha' \simeq \tau'$.

The actual values of τ at a given spot could be significantly greater or less than the regional value, and at a local scale, G and T cannot be neglected.

Let us summarize two of the implications of the longitudinal stress equation.

1. T and $2G$ represent internal resistance to contraction, extension, and flexing of the sheet. These longitudinal stress terms may be responsible for contraction and extension faults, parallel and kink folds, and steeply dipping stylolites, all well-known structures within thrust sheets.

2. T and $2G$ are associated with local variations in surface and basal slope or thickness of the thrust sheet. Over several wavelengths, T and $2G$ cancel out, but on a local scale with short wavelengths they may be the dominant terms. This local longitudinal force $(T + 2G)$ serves to maintain the entire thrust sheet in a continuous creeping movement.

The restrictions deduced so far on the use of (4) can be summarized in points:

1. Equation (4) applies to the deeper portions of thrust faults $H > 5$ km, where $\tau_m/\rho g H < 1$. It is interesting that above such depths the basal slope also steepens rapidly and at the surface may reach 60°. Usually it is only at depths $H > 5$ km that $\beta < 15°$. Equation (4) clearly does not apply to the toe portion of a thrust.

2. The base of the thrust sheet undergoes a large component of simple shear. This may be tested by finite strain analysis.

3. $\bar{\tau}$ and $\bar{\alpha}$ represent regional averages over a longitudinal distance which is several times the thickness H.

In the next section we will apply (4) to a foreland thrust belt.

BASAL SOLE THRUST IN THE CANADIAN ROCKIES

The central and southern Canadian Rocky Mountains constitute the best-known portion of a foreland thrust belt which extends from Mexico to the Arctic Ocean. It is a particularly important area because of the series of rules developed here for interpreting the geometry and succession of thrust faults [e.g., *Dahlstrom*, 1970]. Several carefully constructed 'balanced' cross sections have been drawn through the Canadian Rockies, and we shall use cross sections drawn by *Bally et al.* [1966] and by *Price and Mountjoy* [1970].

The cross sections (Figures 4 and 5) show a large mass of sedimentary rock telescoped together and moved out onto the stable craton. The current lengths of the central and southern cross sections are 208 and 150 km; their original lengths were about 478 and 311 km, indicating a shortening of about 270 and 161 km, respectively, or a compressive natural strain of 83% and 73%. This deformation occurred from the Late Jurassic (Oxfordian) to the mid-Paleocene, a maximum time interval of about 100 m.y. Therefore the minimum mean velocity of shortening is 7×10^{-11} m/s and 5×10^{-11} m/s for the central and southern rockies.

Detailed high-resolution seismic sections, such as those published by *Bally et al.* [1966], reveal that the Precambrian crystalline basement is remarkably smooth, planar, and undeformed. Hence the shortening of about 200 km was restricted to the cover, and there is a major décollement or basal sole thrust separating the rigid crystalline basement from its telescoped cover. All other thrusts in the cover can be thought of as imbricates to this basal sole thrust.

The basement has a gentle dip of 2.5°–3.5° toward the hinterland. The basement did not slope towards the craton at the time of thrusting, as required by gravity gliding theories. In fact the thrusts have moved up the basement slope onto the craton; *Price* [1971] has presented a clear review of this geological evidence.

In the Rockies and other foreland thrust belts the thrusts and major normal faults begin at the basal décollement with very gentle dips and then curve upward with increasing dip (Figures 5 and 6). This distinctive upward-curving shape is known as 'listric' (term discussed in work by *Dennis* [1967, p. 104]). As a consequence of this listric shape, movement causes the curved wedges of rock to overlap each other like shingles in the typical 'imbricate' arrangement. Another typical geometric feature of these faults is the staircase shape of the surfaces, with long flat 'treads' in the incompetent formations and short steep 'risers' through the competent formations (Figures 5 and 6).

What is the current regional shearing stress acting along the basal sole thrust? If we choose a length of about 60 km over which to calculate average values of $\bar{H}$ and $\bar{\alpha}$, then this averaging length is $x_1 > 5\lambda$, and typically the ratio is $H/x_1 \sim 1/10$. Over each 60-km segment of the cross section the average surface slope $\bar{\alpha}$ and the mean thickness $\bar{H}$ can be determined (Figures 4*b*, 4*c*, 5*b*, and 5*c*). By using a typical density of 2.70 $\times$ 10^3 kg m^{-3} the basal shearing stress $\bar{\tau}$ may be calculated from (4) as $\bar{\tau} \simeq \rho g H \bar{\alpha}$.

These regional values of the basal shear stress are shown in Figure 4*d* and Figure 5*d*. Several conclusions can be deduced immediately from inspection of these graphs.

1. A typical order-of-magnitude value for the average basal shearing stress is 2×10^6 Pa (20 bars), possibly with a maximum value of 7.5×10^6 Pa (75 bars).

2. There is a zone of reversed sense of shear under the Main Ranges.

3. On a separate cross section (Figures 4*e* and 5*e*) the magnitude and sense of the shear stress have been indicated by arrows, together with all of the major listric normal faults. These normal faults are slump-type faults of large displacements which have the same general shape as thrust faults (see review in work by *Dahlstrom* [1970]). Listric normal faults are asymptotic to the basal sole thrust and require a specific sense of shear on the sole. There is a striking association: all the major listric normal faults occur over the zone of reversed sense of basal shear determined by (4).

But the Rockies thrust belt does not appear to be active now, and we have used the current surface angle $\bar{\alpha}$ and thick-

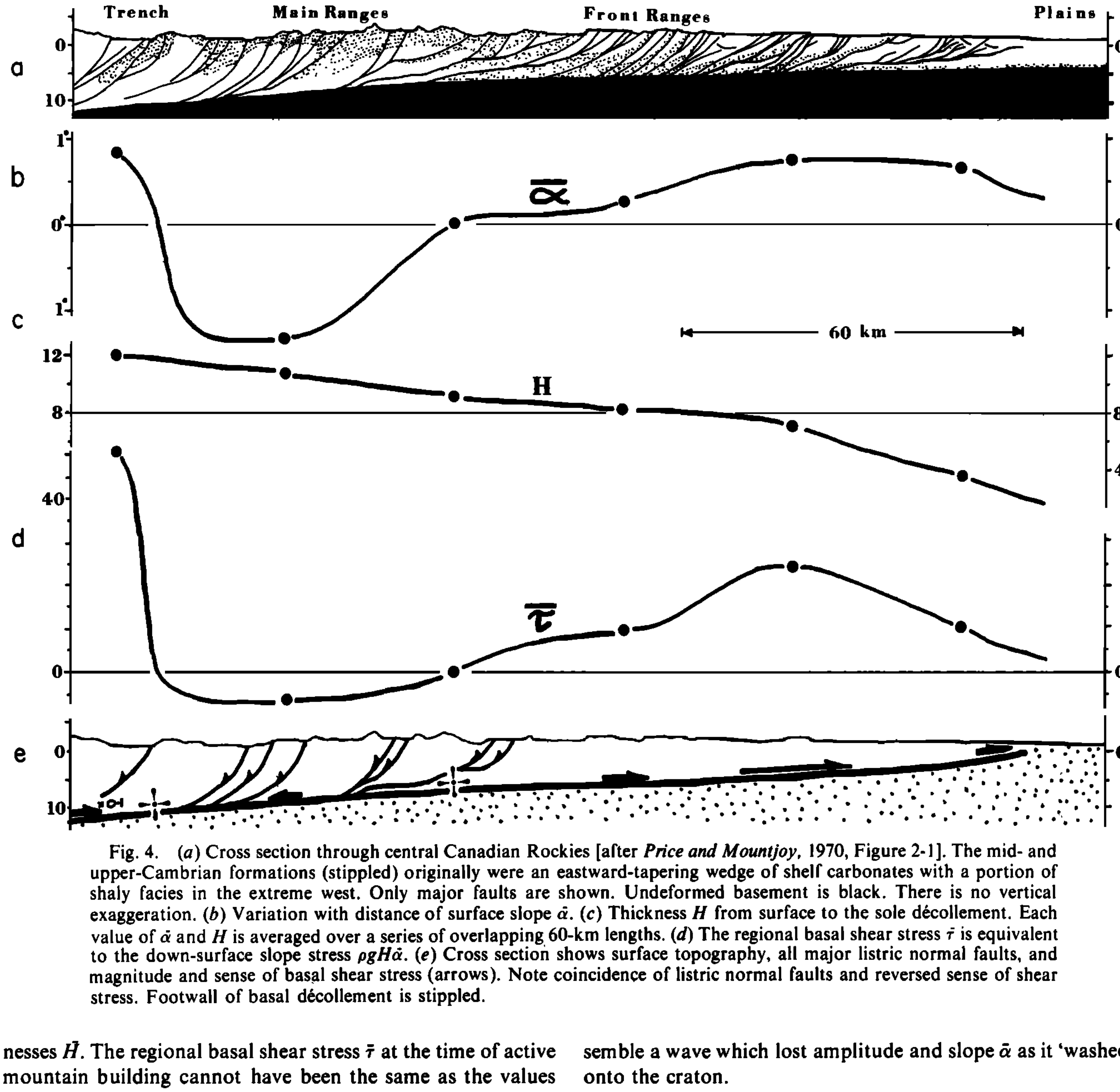

Fig. 4. (*a*) Cross section through central Canadian Rockies [after *Price and Mountjoy,* 1970, Figure 2-1]. The mid- and upper-Cambrian formations (stippled) originally were an eastward-tapering wedge of shelf carbonates with a portion of shaly facies in the extreme west. Only major faults are shown. Undeformed basement is black. There is no vertical exaggeration. (*b*) Variation with distance of surface slope $\bar{\alpha}$. (*c*) Thickness H from surface to the sole décollement. Each value of $\bar{\alpha}$ and H is averaged over a series of overlapping 60-km lengths. (*d*) The regional basal shear stress $\bar{\tau}$ is equivalent to the down-surface slope stress $\rho gH\bar{\alpha}$. (*e*) Cross section shows surface topography, all major listric normal faults, and magnitude and sense of basal shear stress (arrows). Note coincidence of listric normal faults and reversed sense of shear stress. Footwall of basal décollement is stippled.

nesses $\bar{H}$. The regional basal shear stress $\bar{\tau}$ at the time of active mountain building cannot have been the same as the values which we have calculated.

The dip of the crystalline basement of the Atlantic continental margin is from about 0.35° to 0.75°, whereas in the Rockies it is from 2.5° to 3.5°. Possibly the preorogenic continental sediment wedge in the Rockies lay on a basement with the same average dip as the stable Atlantic continental margin. If so, this suggests that at the onset of deformation in the Cordillera the surface sloped an additional 2.15°–2.75° above current values; then if the regional surface slope were slowly reduced, the thrusting would eventually have stopped. There are several ways of reducing the surface slope:

1. Since the end of thrusting in the Paleocene there has been uplift and tilting on a very regional scale of the entire Canadian Rockies (reviewed by *Shaw* [1970]). Possibly a small tilt or change in the very large-scale surface slope would be sufficient to turn off active thrusting.

2. Active erosion during the thrusting may have reduced the surface slope.

3. If the hinterland ceased rising in relation to the craton, the thrust belt would advance onto the craton until it ran out of potential energy. The front of the thrust belt would resemble a wave which lost amplitude and slope $\bar{\alpha}$ as it 'washed' onto the craton.

Probably all these processes have occurred. If the current surface slope is increased by 2°, an order-of-magnitude basal shear stress $\bar{\tau}$ would be 5×10^6 Pa (50 bars). The overall shape of the graph and the gradients derived would be unchanged.

Driving Forces for Thrust Movement

What kinds of forces are responsible for driving thrust sheets? Are they horizontally directed surface forces, caused by a 'push' of the hinterland against the foreland cover rocks, or do thrust sheets move essentially under gravitational forces, driven by the down-surface slope component as we assumed earlier?

The horizontal push theory is usually implied by such statements as 'compressive crustal stresses.' This theory requires that longitudinal surface force must be capable of causing motion at some time or another over the entire length of the thrust sheet. In other words, surface forces applied by the hinterland (left boundary of Figure 6) must be transmitted through the interior of the thrust sheet to its extremities, so that the thrust sheet acts as a 'stress guide.'

If thrusts do not have sufficient strength to act as a stress guide, a push on the boundary would be dissipated or damped before it could accomplish any useful movement. If a sheet shortens easily, then a particular push which moves the trailing edge will not necessarily result in forward motion of the leading edge.

There is a long history of individuals who for various reasons have felt that horizontal surface forces could not move thrust nappes. First, there is an impressive array of field evidence attesting to the overall weakness of thrust sheets (reviewed by *Lemoine* [1973]).

Fifty years ago the Alpine geologists Haarman and Ampferer were, on the basis of such field evidence,

> not inclined to believe that tangential compression could move and over-thrust such bodies as nappes which are not only extremely thin, but also very extensive in horizontal direction. They supposed instead that gravity, affecting every mass element of a tectonic body, would be the only force able to produce such structures

[*Rutten*, 1969].

The inability of horizontal push to move thin, long, thrust sheets might have been anticipated to some extent by dimensional analysis and scale model considerations. Slabs of rock of the order of 10 km thick are mechanically analogous to some semiliquid such as custard in a slab 10–20 mm high [*Hubbert*, 1937]. A horizontal compression on a thrust sheet of large areal extent would be like trying to push a patch of such custard along a board. It will eventually move, but only when one eventually builds up sufficient surface slope.

To examine in a more analytical fashion the horizontal push or longitudinal compressive force theory, we chose a specific stress system. Overlay a rectangle of length L and height H onto the cross section of a thrust sheet, inclined parallel to the surface (Figure 6). A uniform compression C on one end ($x_1 = 0$) of the rectangle, as if it were imposed by a piston travelling subhorizontally, sets up a simple stress system. The normal stress σ_{11}^s set up by this compression C of the rectangle dies out at the other end, $\sigma_{11}^s = 0$ at $x_1 = L$. The normal stress is balanced by a shearing stress σ_{12}^s along the base of the rectangle. The surface is stress free. These supplementary stresses σ_{ij}^s set up by the subhorizontal push are, in the absence of gravity,

$$\begin{aligned} \sigma_{11}^s &= -C + \frac{C}{L}x_1 \\ \sigma_{22}^s &= 0 \\ \sigma_{12}^s &= \frac{C}{L}x_2 \end{aligned} \tag{7}$$

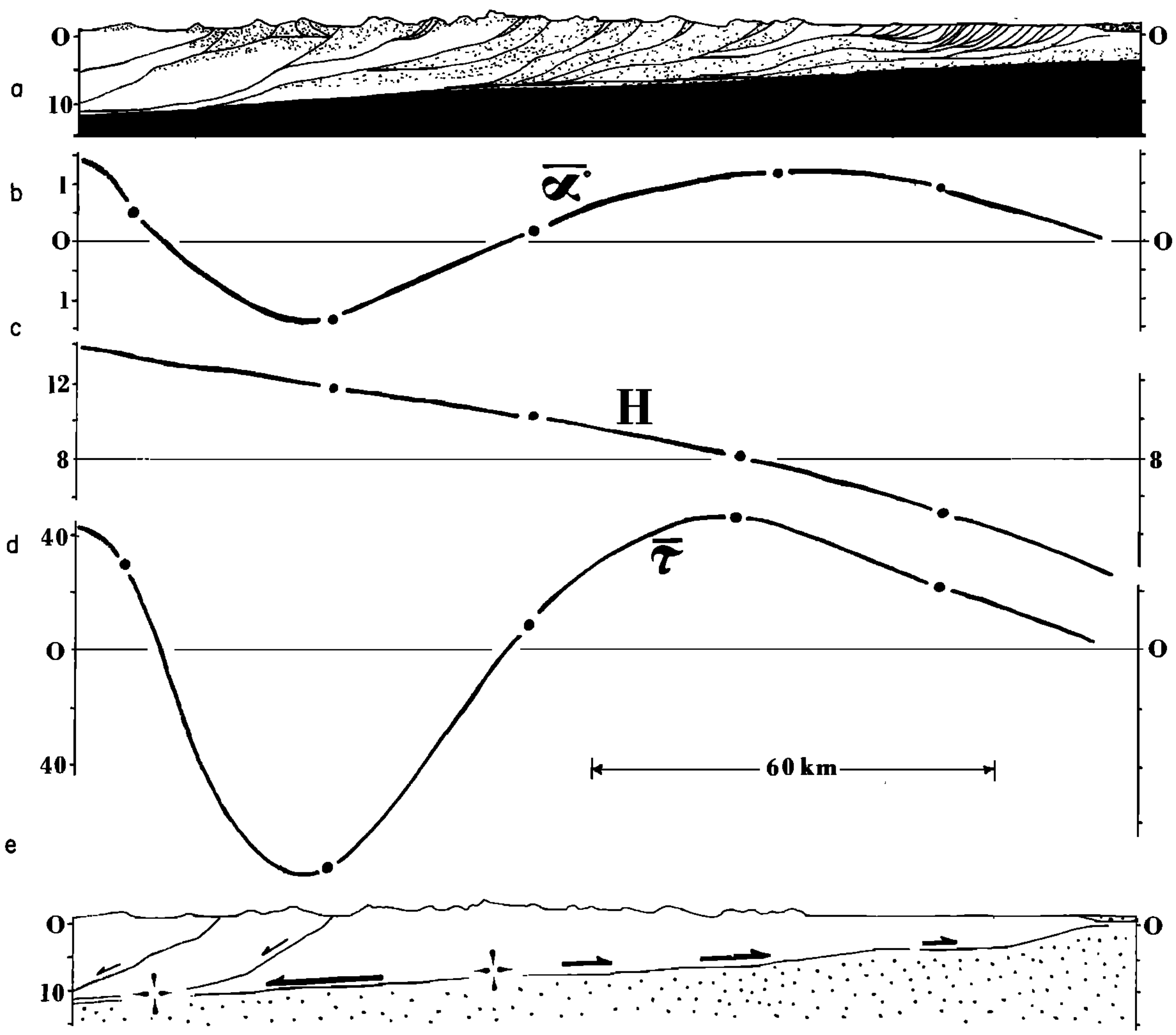

Fig. 5. (*a*) Cross section through southern Canadian Rockies [after *Bally et al.*, 1966, plate 12]. The Paleozoic shelf carbonate wedge has been stippled, and only the major faults are shown. (*b*) Variations in regional surface slope $\bar{\alpha}$. (*c*) Average thickness H. (*d*) Regional basal shear stress $\bar{\tau}$. (*e*) Relations between $\bar{\tau}$ and listric normal faults calculated as in Figure 4.

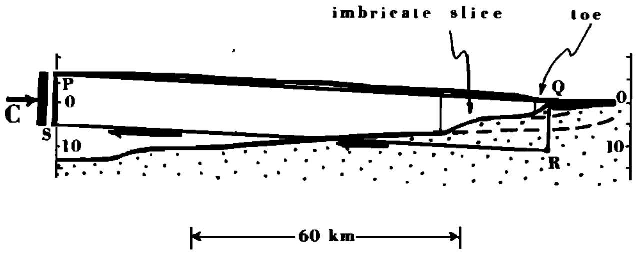

Fig. 6. Typical thrust, based on the McConnell thrust at the onset of formation [e.g., *Bally et al.*, 1966, plate 12]. Minimum length is $\simeq 85$ km, possible length is $\simeq 250$ km, and average thickness is $\simeq 10$ km, so ratio(L/H) is between 8.5 and 25. Cross-section area of toe is $\simeq 5$ km^2, major imbricate slice is $\simeq 90$ km^2, and remaining main body of sheet is between 1250 and 2600 km^2. Dip of base $\bar{\beta}$ for toe is $\simeq 40°$, major imbricate slice is $\simeq 10°$, and main body is $\simeq 3°$. Assumed average surface slope $\bar{\alpha}$ is $\simeq 3°$. Rectangle PQRS has the mean dimensions of the thrust. Supplementary stresses imposed by surface forces (equation (7)) are a subhorizontal compression C balanced by shearing stress along base RS. PQ approximates ground surface and is stress free. Gravitational stresses (equations (3)) are not shown. Total stress system (equations (8)) has dotted trajectories of maximum shear stress.

This supplementary system of stresses was suggested by *Hafner* [1951], who demonstrated them to be in equilibrium.

The down-surface slope stresses acting alone constitute another stress system in equilibrium. Since the statically determinate horizontal push (equation (7)) and down-surface slope (equations (3*a*) and (3*b*)) stress systems are in equilibrium, they may be added to get a result which is also in equilibrium. After expressing the horizontal push C as a percentage k of the down-surface slope compressive normal stress, so that $k = C/\rho g H \cos \alpha$,

$$\begin{aligned}\sigma_{11} &= -\rho g H \cos \alpha\left(1 + k - \frac{x_1 k}{L}\right)\\ \sigma_{22} &= -\rho g H \cos \alpha \qquad (8)\\ \sigma_{12} &= \rho g H\left(\sin \alpha + \frac{x_2 k}{L}\right)\end{aligned}$$

This new stress system contains the combined effects of a subhorizontal compression and down-surface slope stresses; all the body and surface forces are shown acting on the rectangle of Figure 6.

A special case of this stress system, surface slope $\alpha = 0$, so that there are no down-surface slope stresses, was first proposed for thrust faults by *Hubbert* [1951] and *Hafner* [1951] and clarified further by *Jaeger* [1969]. These authors point out the qualitative resemblance between the system's upward-curving 'listric' shaped trajectories of maximum shear stress and natural thrust faults (Figure 6). We shall see that it is unrealistic to ignore the surface slope.

In an earlier section we examined longitudinal stress gradients from the point of view of (1) the mean normal stress approximating the weight of overburden and (2) an exact longitudinal stress equation which has undergone considerable theoretical development in glaciology. With the advantages of a fully determined state of stress (equation (8)) we can now investigate the longitudinal stresses in a more clear-cut and concrete fashion and compare the gravitational to the compressive surface forces for various geological conditions.

Thrust sheets of large dimensions contain a whole spectrum of contraction faults, or minor thrusts, along which very much less movement occurred than along the major thrusts. Yet because thrust sheets remain reasonably coherent, the long-range average shear stress on all the other surfaces must have remained less than on the main thrust. Otherwise the thrust sheet would collapse into a sequence of discontinuous bodies, and the toe of the thrust would remain stationary.

The rectangle of height H and length L is positioned to fill much of the cross-sectional area of one active thrust (Figure 6). This rectangle is acted on by both the subhorizontal compression and the down-surface slope stresses. It is necessary that the maximum shear stress within the rectangle should reach the value of the rock strength only along the thrust, not within the sheet, so

$$\left(\frac{\sigma_{11} - \sigma_{22}}{2}\right)^2 + \sigma_{12}^2 \leq \tau_m^2$$

Substituting in (8) and noting that the maximum value of shear stress is found along the thrust at a depth H at the end where the compression is applied,

$$\left(\frac{\tau_m}{\rho g H \cos \alpha}\right)^2 = \left(\frac{k}{2}\right)^2 + \left(\tan \alpha + k\frac{H}{L}\right)^2 \qquad (9)$$

A graph of (9) for (H) versus L/H is shown in Figure 7. It is clear that gravitational forces are larger than compressive forces for all parts of thrusts, and for entire major sheets, gravitational forces completely overwhelm any possible longitudinal compressive surface forces.

This result is not a consequence of choosing a particularly simple system of horizontal compressive stresses. The same method applied to all the stress systems for thrusts proposed by *Hafner* [1951] demonstrates that in all these cases k is negligibly small.

It would appear that most thrusts move essentially under the influence of stresses induced by the down-surface slope gravitational component. Compressive stresses set up by the hinterland pushing horizontally against its foreland are negligible in so far as entire thrust sheets are concerned.

We are in agreement with the early workers such as Haarman and Ampferer and their widely accepted 'gravity gliding' theory on some fundamental points. Without any relative topographic relief there is no thrusting, and everything is stable. Events occurring within the hinterland are in a sense 'transmitted' to the foreland as stresses, but these stresses are a

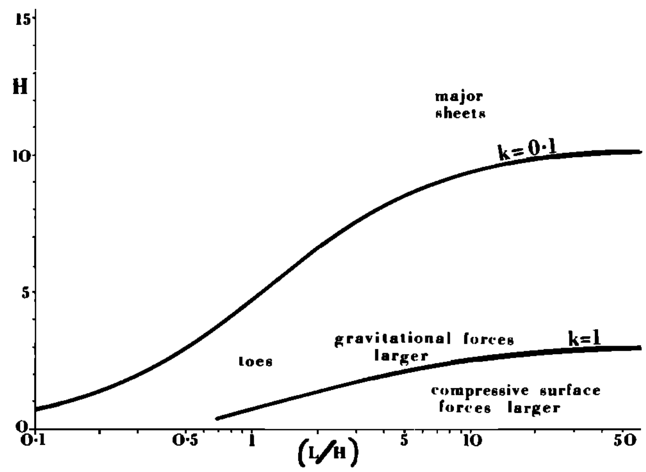

Fig. 7. Graph of H versus (L/H) determined from (9) for $k = 0.1$ and $k = 1$ by using $\bar{\alpha} = 3°$. Entire major thrust sheets have dimensions which place them within the $k < 0.1$ sector, but for toes of thrusts $k \simeq 0.5$, and compressive surface forces are important.

direct consequence of the rise or fall of the hinterland. Let us look at some of the larger-scale consequences of these results.

The time relations between the magmatic activity and thrusting in the North American Cordillera from the Mesozoic to early Eocene were reviewed by *Armstrong* [1974]. He points out that increased activity in the Andean-type magmatic arc was immediately followed by a cratonward surge in the foreland thrust belt. As magmatic activity reached a climax and terminated at different times along the belt, so did the development of the immediately adjacent portion of the foreland thrust belt.

An explanation of this striking association between magmatism and foreland thrusting is given by *Armstrong* [1974] and *Burchfiel and Davis* [1975], which I understand as follows:

1. Plutonism enhanced the ductility of the crust, leading to a lower shear strength τ_m.

2. Horizontal compressive stresses generated at the leading western edge of the North American plate were a dominant force all the way from the magmatic arc as far east as the foreland thrust belt. This compression was able to produce shortening in the ductile crust.

An alternative explanation is that increased ductility and uplift in the vicinity of the magmatic arc lead to an increase in regional surface slope, which in turn sets off thrusting.

On an even larger scale there is a broad consensus that continental margin orogeny and the formation of foreland thrust belts are produced whenever continental collision and suturing occur [e.g., *Dewey and Kidd,* 1974]. *Drewry et al.* [1974] assume that pre-Permian orogenic belts represent continental collisions and use this as a guide in constructing pre-Permian paleogeographic maps. Yet the best example of a currently active continental margin orogen is the Andes, where continental collision did not occur during the Cenozoic nor is it occurring now [*James*, 1971]. *Burchfiel and Davis* [1972] claim that there were no major continent-continent collisions during the development of the American Cordillera.

It is sometimes implied that orogeny is produced by horizontal compressive surface forces arising from continent-continent collision. The analysis presented above suggests that the immediate cause of orogenic folding and thrusting is the development of a sufficient regional surface slope. This slope is determined by uplift in the hinterland region surrounding the magmatic arc, and the uplift in turn is fixed ultimately by the position and consumption rate of the subduction zone. The episodic nature of orogenic deformation is related to the episodic nature of uplift in the magmatic arc. This hypothesis does not exclude continental collision; nor, however, is it necessary solely to produce an orogeny.

POTENTIAL ENERGY OF A THRUST SHEET

To move forward entirely by gravity, a column from the surface to base of thrust (Figure 8) must lower its center of gravity $(H/2)$ as the sheet advances, so

$$\frac{DH}{Dt} \leq 0 \tag{10}$$

This is differentiation with respect to time t following the motion of that particular column. The velocity v is that of the column at position x_1, where x_1 is now horizontal (Figure 8). Therefore

$$\frac{DH}{Dt} = \frac{\partial H}{\partial t} + v\frac{\partial H}{\partial x_1}$$

The height of the column H varies with the amount of erosion or deposition B at the surface and the vertical strain e_{22} of the column, or $H = f(B, x_1, e_{22})$; therefore

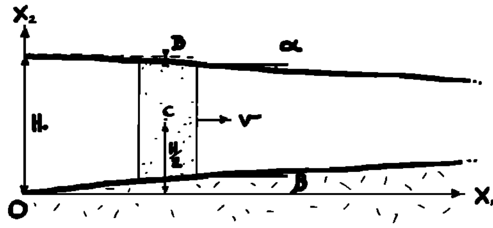

Fig. 8. Portion of thrust sheet with surface and basal slopes α, β is as before, but the Cartesian coordinate frame is now oriented with x_1 horizontal and x_2 vertical. A marked columnar mass (stippled) is at position x_1, has center of gravity C at a vertical height $(H/2)$, and is moving with velocity v. This column started from the origin O where it had a height H_o. In travelling to its current position a thickness β was eroded from the surface of the column.

$$\frac{\partial H}{\partial t} = \frac{\partial H}{\partial B}\cdot\frac{\partial B}{\partial t} + \frac{\partial H}{\partial x_1}\cdot\frac{\partial x_1}{\partial t} + \frac{\partial H}{\partial e_{22}}\cdot\frac{\partial e_{22}}{\partial t}$$

$$\frac{DH}{Dt} = \frac{\partial H}{\partial B}\cdot\frac{\partial B}{\partial t} + \frac{\partial H}{\partial e_{22}}\cdot\frac{\partial e_{22}}{\partial t} + \frac{\partial H}{\partial x_1}\left(v + \frac{\partial x_1}{\partial t}\right) \tag{11}$$

It is well established that thrust sheets of even unmetamorphosed rocks undergo strains as they move forward. For example, the thrust sheet may be thickened by contraction faults or thinned by extension faults. The finite strain measure used here is the natural strain

$$e_{22} = \ln\left(\frac{H}{H_o - B}\right) \qquad \left(\frac{\partial e_{22}}{\partial H}\right)_B = \frac{1}{H}$$

where H_o is the original and H the current length (Figure 8).

The strain rate is $\dot{e}_{22} = \partial e_{22}/\partial t$, and the velocity $v = \partial x_1/\partial t$.

We will assume erosion is in progress, so B is negative. The rate of erosion is $\dot{B} = \partial B/\partial t$. Since B and H are directly linked so that an increment of erosion ΔB gives less height ΔH by the same amount, $\partial H/\partial B = 1$. Also

$$\frac{\partial H}{\partial x_1} = \tan\alpha + \tan\beta$$

Substituting the above relations into (11),

$$\frac{DH}{Dt} = H\dot{e}_{22} - \dot{B} + 2v(\tan\alpha + \tan\beta)$$

and from (10), after rearranging,

$$\tan\alpha \geq \tan\beta - \frac{H\dot{e}_{22}}{2v} - \frac{\dot{B}}{2v} \tag{12}$$

The surface slope α is here dipping in the direction of motion, and the signs of $\dot{B}$ and $\dot{e}_{22}$ are here negative when H is thinned by erosion or contraction strains. This inequality (12) must be obeyed if the columnar mass is to move forward under gravity.

We see immediately that forward motion of a thrust sheet is made easier through (1) erosion $\dot{B}$, (2) deformation $\dot{e}_{22}$, (3) decrease in dip β of the thrust plane, and (4) increase in surface slope α.

TOES, IMBRICATION, AND PIGGYBACK THRUSTING

After following a ductile horizon for some distance a thrust may curve smoothly upward to reach the topographic surface at a steep dip (e.g., Figures 4 and 5). Erosion is intense at the leading edge, and much debris is produced. Continual movement of the thrust requires that it cut through and travel over its own detritus. This curved wedge making up the frontal portion of a thrust sheet is known as the toe. A typical cross-sectional area of a toe would be 5 km².

Because of the steep basal slope, often 30° or more, the simple equation (4) cannot be used. Further, as the toe advances, its center of gravity may rise, and forward motion cannot be accomplished by gravitational forces operating on the volume of the toe alone. The thrust sheet as a whole must lose potential energy, but the toe portion gains some. One cannot move the toe without a horizontal push, but the horizontal push on the small toe volume is accomplished by the gravitationally driven main body of the thrust sheet.

We saw earlier that imbrication is a geometric consequence of motion on listric thrust surfaces. It can be found on a small scale, making up part of a toe, but on a larger scale, all thrust slices observed at the surface may be imbrications from a basal décollement. The cross-sectional area of a major imbricate slice is typically 90 km², measured above and in front of the junction with the basal décollement. The dip of an upward-curving imbricate thrust is roughly 10° (Figure 6).

New imbricate thrusts cut under the older ones, and one result of this successive imbrication of the footwall is that material is removed from the footwall and accreted onto the hanging wall. As a new imbrication moves forward, it carries the older thrust along piggyback. This piggyback thrusting sequence has been known from the western Alps for at least 75 yr and is the common order in the Canadian Rockies [e.g., *Dahlstrom*, 1970]. The effect of surface slope in producing piggyback stacks can be described qualitatively. Initially, the thrust belt stands off some distance from some area of interest which has a comparatively horizontal topography and low surface elevation. Eventually, the mountainous thrust belt advances to one side of the area of interest so that the entire region acquires a long-range surface slope towards the craton. The new surface slope sets up a basal shearing stress, which eventually becomes large enough along some potential décollement horizon to set a new thrust sheet into motion. This new thrust underlies the older ones.

Although the regional paleoslope must be in the direction of thrusting, at least over the initial site of the thrust sheet, it is not necessary to maintain this paleoslope simultaneously over the entire thrust belt. Haarman's 'surfrider' analogy [see *Lemoine*, 1973] is appropriate, but now we see that an advancing wave produces a surface slope which drives thrusts ('surfboards') ahead of it by the down-surface slope stresses.

It follows that piggyback thrusts can be used to deduce the evolution of the paleoslope with a law reminiscent of Walther's law of facies. Higher thrust sheets have travelled further than lower ones, and at the onset of a particular imbrication the paleosurface dip direction was in the same sense as motion on the imbrication. The vertical structural succession gives the horizontal evolution.

The paleoslope deduced from a piggyback stack of thrust sheets may be exactly opposite in sense from the paleoslope determined from study of the sediments making up the thrust sheets. In other words, the deeper water sediments often make up the higher thrust sheets, implying a large relief inversion between sediment deposition at a stable Atlantic-type continental margin and the onset of subsequent thrusting.

There are good field criteria for proving whether or not a particular stack of thrusts developed in piggyback fashion, quite independent of any physical theory of how they formed:

1. Piggyback stacks always have a systematic order to the tectonic superposition. Even though an upper thrust sheet may directly overlie any of the lower sheets, the vertical succession of thrust sheets is never reversed.

2. Active thrust fault terrain produces a clastic wedge whose source area is the moving thrusts themselves. Individual thrusts may produce a characteristic detritus. This detritus accumulates in front of the emerging toe of the advancing sheet and is overriden by the thrust shortly after deposition (e.g., the Aruma in Figure 12).

These synorogenic sediments provide information not only on the piggyback movement of imbricate stacks but also give independent sedimentological evidence of the regional surface slope.

Thrust sheets consisting of the old continental shelf and slope sediments will feed a siliceous and calcareous synorogenic clastic wedge (Molasse). These thrusts often cut up to the surface above sea level, and sediment accumulates as coalescing alluvial fans at the margin of a growing mountain range. Study of the synorogenic sediment wedge in the Rockies clearly demonstrates that the thrust belt was high and mountainous in comparison with the adjacent undeformed foreland, that the thrust belt advanced towards the craton, and that the Molasse was deposited in intercratonic foredeeps immediately ahead of the advancing thrust sheets (reviewed in work by *Bally et al.* [1966] and *Price and Mountjoy* [1970]). This is also true in the western Alps [*Trümpy*, 1960].

EMPLACEMENT OF OPHIOLITE THRUST SHEETS ON CONTINENTAL MARGINS

Because oceanic lithosphere is denser and has a lower surface elevation than continental lithosphere, one does not expect to find oceanic lithosphere on top of continental. Yet ophiolite slabs of very large dimensions rest upon unquestionable continental crust in some of the most spectacular thrust nappes. Those found in the Oman, New Guinea (Papua), and western Newfoundland are made up of oceanic crust and mantle in enormous gently dipping sheets, of the order of 10 km thick and up to 100 km in cross-section length. They are not rare; in fact, the emplacement of such ophiolites may be one of the characteristic mountain-building processes. Yet 'the mechanics of displacing oceanic crust and mantle onto a continental margin are either completely unknown, or at most, poorly understood' [*Williams and Smyth*, 1973, p. 616]. This paradoxical process whereby oceanic crust and mantle is thrust over continental crust has been called 'obduction' by *Coleman* [1971], and a wide variety of mechanisms have been proposed. In this section I wish to reappraise the emplacement of ophiolites in terms of the theory that we have developed. Two hypotheses are particularly popular, of which there are several minor variations, and they are outlined below and sketched in Figures 9 and 10.

The first model involves thrusting of ophiolite slabs from the leading edge of the overriding oceanic plate during the attempted subduction of a continent [*Moores*, 1970; *Dewey and Bird*, 1971; *Coleman*, 1971; *Oxburgh*, 1972]. The thrust faults at the base of ophiolite complexes would be the subduction zone itself according to this interpretation. The upper portion of oceanic lithosphere 'splinters' or 'flakes' off and is driven up and over the continent (Figure 9). The Papuan ophiolite nappe [*Davies and Smith*, 1971] and the Semail ophiolite nappe in the Oman [*Dewey et al.*, 1973] have been proposed as examples of this process.

The second model starts with development of thrust slices within the oceanic lithosphere attached to a continent [*Rabinowitz and Ryan*, 1970; *Dewey and Bird*, 1971]. These slices form

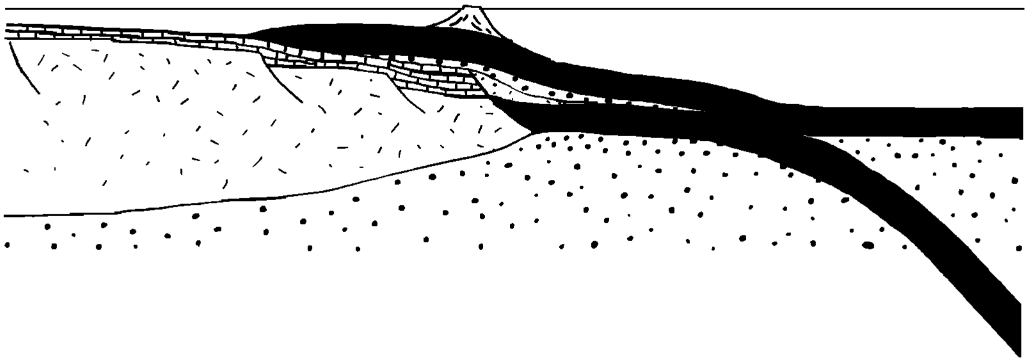

Fig. 9. Continent on downgoing slab reaches subduction zone; then oceanic crust (black) and some mantle (large stipple) and island arc assemblage of the overriding plate are thrust up and over continental rise and slope sediments (fine stipple) onto the continental shelf as a large 'flake.'

a growing ridge on the downgoing slab, and as the continent approaches subduction, the ophiolite thrust sheets are moved back up and over the continent (Figure 10). It has been suggested that the ophiolites found in western Newfoundland [*Dewey and Bird*, 1971] and again in Papua [*Milsom*, 1973] were emplaced in this manner.

The major difference between the two hypotheses is whether the ophiolites originate from the overriding (Figure 9) or the downgoing plate (Figures 9 and 10). Both hypotheses assume that the continent is part of a downgoing slab which has reached the vicinity of a subduction zone. The obduction results from attempted subduction of the buoyant continental margin. Because it is impossible for a buoyant continent to move very far down a subduction zone, it is claimed that after ophiolite emplacement the subduction zone either undergoes a 'flip' in direction or continually shifts its position away from the continental margin [*McKenzie*, 1969; *Dewey and Bird*, 1971; *Dewey et al.*, 1973]. One objection to both these mechanisms is that nowhere today can one find a continent on the verge of going down a subduction zone. In all cases the subduction zones dip beneath the continent.

Both mechanisms imply that ophiolite thrust sheets are pushed out by horizontal surface forces. These forces are supposed to move thrust sheets 'up and over' the continental crust and are frequently illustrated moving the thrust sheets up against the overall surface dip. These compressive surface forces are supposed to arise during continent-continent collision.

But we shall see that there are strong similarities between ophiolite thrust sheets and the foreland thrusts which involve shelf sediments. Let us first review a particularly informative thrust belt involving ophiolites recently described from the Oman, where *Glennie et al.* [1973] worked out the following sequence of events:

1. There is the mid-Permian to mid-Cretaceous development of an Atlantic-type continental margin, consisting of shelf carbonates (Hajar supergroup) with facies changes offshelf out to the slope and rise sediments (Hawasina formations) and then oceanic crust. The width of this ocean from shelf to spreading center in mid-Cretaceous cannot have been less than 400 km and may have been more than 1200 km (Figure 11).

2. In Late Cretaceous the relief inverted, and a basin formed on the shelf. On the oceanward side of this basin the shallow water (Hajar) carbonates were uplifted and shed into the basin a conglomerate which fined into shales toward the craton (Aruma group). Some of the conglomerate contains ophiolite debris, indicating presence of a nearby ophiolite sheet which must have stood higher than the intracratonic basin.

3. Sedimentation in this intracratonic basin ceased abruptly with arrival of a piggyback stack of thrust sheets from the site of the former ocean (Figure 12). The lower 12 sheets (Hawasina unit) of this imbricate piggyback stack consist successively upwards of sediments from the slope and rise. On top of these sheets of slope and rise sediments is one huge ophiolite thrust sheet (Semail nappe) with a current outcrop width of 100 km and thickness before erosion of 10 km. This stack (Hawasina and Semail) travelled over the continental margin to arrive at its present position, a distance of at least 120 km. It is important to note that continent-continent collision did not occur during emplacement of the Oman ophiolites [*Coleman and Irwin*, 1974].

Ophiolite thrust sheets, such as in the Oman, give rise to

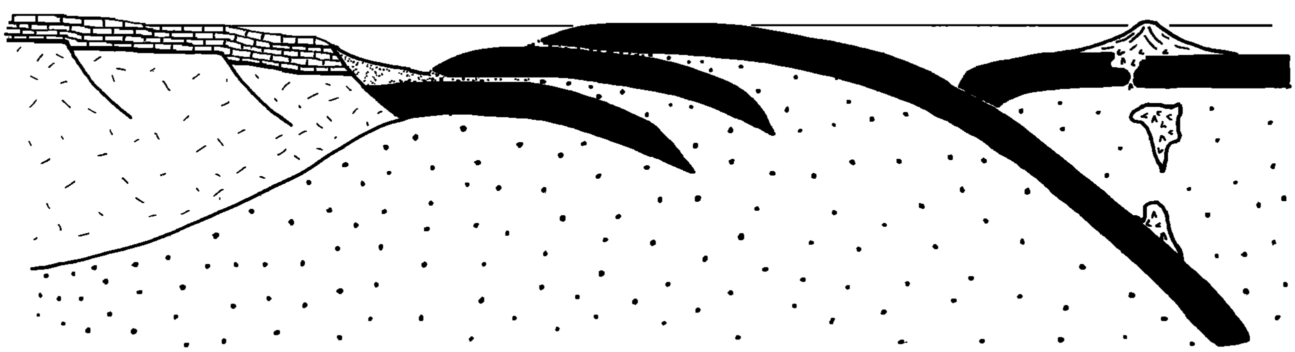

Fig. 10. Thrust sheets of oceanic crust and mantle from a ridge left of the downgoing slab. These thrust sheets are about to move over continental slope and rise sediments. Island arc igneous rock occurs in overriding slab (V pattern).

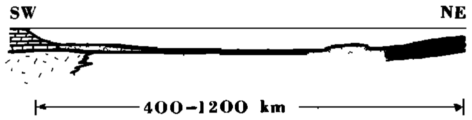

Fig. 11. Palinspastic sketch of mid-Cretaceous depositional environment in the Oman [after *Glennie et al.*, 1973, Figure 8*b*]. The continental margin at SW grades into slope, rise, and oceanic sediments making up various formations in the Hawasina (stippled). Oceanic crust and mantle (black) of the Semail ophiolite lie an unknown distance to SW of a spreading center. Width of this part of the mid-Cretaceous Tethys Ocean was at least 400 km and possibly 1200 km.

particular kinds of clasts eroded from various levels within the ophiolite, deposited in front and then overridden by the advancing thrusts. Ophiolite detritus showing these relations is well developed below the Ligurian ophiolite sheet and has been called 'precursor olistostrome' [*Elter and Trevisan*, 1973].

Ophiolite thrusts may cut up to the sea floor. Erosion is probably submarine; transport is via debris flows and turbidity currents, and deposition is in fairly deep water. A fine matrix could be deep sea clays and silts stirred up by current activity surrounding the advancing thrusts and associated submarine fans. Consequently, one would expect a synorogenic sediment deposited in front of advancing ophiolite thrust sheets to show the following features: unsorted exotic debris from various levels within the ophiolite transported by and mixed with turbidites and debris flows, all deformed by the overriding thrust sheet.

In western Newfoundland a piggyback structural succession of thrust sheets is observed similar to the Oman [*Williams et al.*, 1972; *Williams*, 1973]. Paleogeographic reconstruction of the western Newfoundland ophiolites suggests a minimum length of thrust surface of about 100 km [*Williams et al.*, 1973]. Analysis of gravity and magnetic data suggests that the minimum length of the thrust beneath the Papuan ophiolite is 90 km and it is about 15 km thick [*Milsom*, 1973].

Like thrust sheets of shelf rocks, (1) ophiolitic sheets have erosion debris accumulating in intracratonic troughs ahead of the advancing thrusts, (2) they occur in imbricate stacks which have moved piggyback, (3) they are substantially intact, and (4) they are so very long and thin that ophiolite thrust sheets must move entirely under gravitational forces in the direction of surface slope (see Figure 7).

If it is true that piggyback thrust sequences involving ophiolites are driven essentially by the down-surface slope stresses, this implies a regional surface sloping and moving toward the continent, picking up and driving in turn the ophiolite, rise, slope, and possibly shelf thrust sheets. A trough on the continental margin accumulating debris from the advancing ophiolite thrust sheets, such as observed from the Late Cretaceous in the Oman, demonstrates that this surface slope existed. The formation of such a trough would also help produce the regional surface slope.

A candidate for producing the rest of the regional surface slope might be a rising island arc, separated from the continent by a marginal basin floored by oceanic lithosphere. If such an island arc were rising at a rapid enough rate, the required surface slope toward the continental margin would initiate thrust sheets directed toward the continent.

An example in the Cordillera of Nevada and California is reviewed by *Buchfiel and Davis* [1972, 1975]. The Roberts Mountain thrust sheet of oceanic and slope sediments was emplaced upon the continental crust during Late Devonian-Early Mississippian partial closing of a back arc basin between the continental margin and an offshore Klamath-Sierran arc. This back arc basin finished closing in Early Triassic with the emplacement of the Golconda thrust sheet of oceanic rocks. *Buchfiel and Davis* [1972, 1975] point out that emplacement of these two major thrust sheets coincided with periods of intense andesitic volcanism in the Klamath-Sierran arc. Is it possible that the emplacement of these oceanic sheets also corresponds with intense uplift in the magmatic arc?

There seems little doubt that magmatic arcs are among the most rapidly and continuously uplifted of all tectonic features. For example, evidence of such uplift is provided by Quaternary raised terraces and beaches and some geodetic data from the eastern Aleutians and the Andes [*Plafker*, 1972]. The Mesozoic–early Tertiary Klamath-Sierran magmatic arc underwent extremely rapid uplift and erosion and during its active life was a principal source of sediments to the surrounding basins [*Buchfiel and Davis*, 1972]. On the basis of evidence from deep-penetration reflection seismic lines shot for hydrocarbon exploration, *Beck and Lehner* [1974] claim that magmatic arcs in general show strong continuing uplift and further that there is collapse and subsidence of entire marginal basins between arc and continent. Such processes should produce thrust sheets, but I have not seen any published reflection seismic lines through such modern marginal basins.

In short, it is claimed here that ophiolites are emplaced onto continental margins by gravitational down-surface slope stresses. This implies a specific paleoslope at the time of emplacement but does not imply any particular polarity of the subduction zone, nor does it require continent-continent collision.

Arc-Trench Gap as a Thrust Belt

The steepest regional surface slopes within an orogen are probably found in the arc-trench gap. Consequently, one should find here a high density of thrusts dipping toward the magmatic arc. Deep-penetration reflection seismic lines across the Java trench show an imbricate stack of thrust sheets emerging at the inner trench wall [*Beck and Lehner*, 1974]. Seismic lines across other arc-trench gaps show that this is a general situation [*Hamilton*, 1974; *Seely et al.*, 1974]. These imbricate thrusts are asymptotic to a basal décollement which in this case is the Benioff zone itself. Magmatic arcs situated on continental margins, such as the east Aleutians and the Andes, also appear to have imbricate thrusts asymptotic to the Benioff zone as décollement [*Plafker*, 1972]. I for one am convinced that these imbricate thrusts are typical of arc-trench gaps.

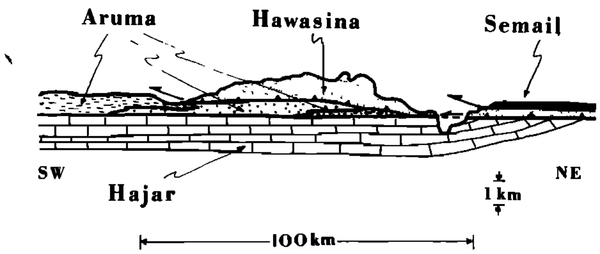

Fig. 12. Cross section through Oman thrust belt, based upon *Glennie et al.* [1974, Figure 2], but here effects of postthrusting Tertiary deformation are removed and restored to angular unconformity above Hajar supergroup of shelf carbonates. The Aruma group of synorogenic clastics grades from coarse conglomerates in center out toward shales at SW. The Hawasina (fine stipple) is a complex of 12 thrust sheets of which only the basal one is shown and is overlain by the Semail ophiolite thrust sheet (black), which along this line of section is deeply eroded.

Trenchant criticism of views to the contrary, usually based on shallow-penetration sparker profiles with from 40 to 100 times vertical exaggeration, has been presented by *Plafker* [1972, p. 919].

A fossil example is Mesozoic–early Tertiary California, where the Great Valley and Franciscan units accumulated in the arc-trench gap, and their structure is dominated by thrusts dipping towards the Klamath-Sierran volcanic arc [*Ernst*, 1970]. Thus on a larger scale the Mesozoic–early Tertiary Cordillera thrusting was two sided, thrusts in both foreland thrust belt and arc-trench gap dipping toward the intervening magmatic arc [*Buchfiel and Davis*, 1972, 1975]. Is it possible that both of these thrust belts arise in the same way?

In describing the development of a foreland thrust belt a useful reference point is one chosen ahead of the thrust belt and on the basement. For example, from a fixed point on the exposed Precambrian craton the Canadian Rocky Mountain thrust belt and synorogenic foredeep would advance toward the reference point at a velocity of about 5×10^{-11} m/s. If the reference point for some reason were chosen close to the thrust belt, the point would travel on the basement but remain immediately below the décollement.

Describing the arc-trench thrust belt in a similar fashion, we choose a reference point on the oceanic crust some distance in front of the outer trench slope. From this reference point the thrust belt would advance towards us at an observed rate of 2.5×10^{-10} m/s to 2.5×10^{-9} m/s (10–100 mm yr^{-1}.

Is it possible that the arc-trench thrust belt advances toward the reference point by initiating thrusts from the base of the regional slope (i.e., base of the trench) and thus 'accreting' thrust sheets of sediment and oceanic lithosphere in a piggyback stack fashion? Could some of the sediment accumulating in the trench be erosion debris from these active thrusts, (analogous with Molasse foredeep or foredeeps in front of ophiolite sheets)?

Is the main driving force for the thrusts in the arc-trench gap compression and drag from the downgoing slab, or could these thrusts be driven by the down-surface slope stress, with the regional surface slope being produced by uplift and igneous activity at the magmatic arc?

CONCLUSIONS

1. There is a simple equation for the regional average basal shear stress $\bar{\tau}$ along a thrust which is deeper than about 5 km. It is $\bar{\tau} \simeq \rho g H \bar{\alpha}$.

2. Thrusts always move in the direction of the topographic surface slope $\bar{\alpha}$, even if this means moving 'up' the basal slope.

3. The basal sole thrust in the Canadian Rockies foreland thrust belt had a regional basal shear stress of the order of 5×10^6 Pa (50 bars).

4. The major listric normal faults in the Rockies coincided with a zone of reversed sense of basal shear stress beneath the main ranges.

5. Whether horizontal tectonic or gravitational forces dominate in the formation of a particular thrust depends upon the magnitude of the dimensionless number k which is related to α, H, L, and τ_m by

$$\left(\frac{\tau_m}{\rho g H \cos \alpha}\right)^2 = \left(\frac{k}{2}\right)^2 + \left(\tan \alpha + k \frac{H}{L}\right)^2$$

6. For thrusts as a whole this dimensionless number, k ($=C/\rho g H \cos \alpha$), is far less than 1, and gravitational forces dominate. Toes advance under compressive forces imposed by the main body of the thrust sheet, which itself is gravitationally driven.

7. Along the Mesozoic–early Tertiary North American Cordillera the close association between activity in the magmatic arc and development of the adjacent foreland thrust belt may be a consequence of increased surface slope produced by processes associated with the magmatic arc.

8. Neither foreland thrust belts nor ophiolite thrust sheets necessarily imply or require continent-continent collision for their development.

9. A necessary restriction on moving thrusts by gravitational potential energy is that

$$\tan \alpha \geq \tan \beta - \frac{H\dot{e}_{22}}{2v} - \frac{B}{2v}$$

10. Imbrication and piggyback thrust stacks are also a consequence of the paleoslope and may be used to determine it.

11. Thrust sheets of oceanic crust and mantle were emplaced upon continental crust by gravitational forces, not horizontal compressive surface forces. 'Obduction' is not substantially different from any other kind of thrusting; the material being moved is different, but the physical processes appear similar.

12. The thrust belt in the early stages of the orogeny might in some cases have stretched from the craton, across a back arc marginal basin, to the andesitic volcanic arc. Uplift of the andesitic volcanic arc might result in a piggybacked stack of oceanic thrust sheets on a continental crust. This could explain the Paleozoic-Mesozoic Roberts Mountain and Golconda thrust complex in the Nevada-California Cordillera.

13. One possible cause of the thrusts in the arc-trench gap is the extremely steep regional surface slope in this tectonic province. This speculation should be investigated, and the down-surface slope stresses compared with the drag produced by the downgoing slab.

This paper has been concerned largely with the regional implications of the mechanics of thrust faulting. What has been outlined followed mainly from static considerations and is to a large degree independent of the material properties. We have avoided the question of what physical mechanisms actually operate along the thrust fault surface during slip, whether, for example, a solid friction law with pore pressure is obeyed, or whether sliding proceeds by some other process such as viscous flow.

However, there is a high density of minor structures along thrust faults and within thrust sheets, such as stylolites, contraction and extension faults, kink folds, and slip on most bedding surfaces. When they are combined with energy considerations, these minor structures provide a different kind of information on the mechanics of thrusting. This will be discussed elsewhere [*Elliott*, 1973*a*, 1975].

NOTATION

- ρ density, equal to 2.7×10^3 kg m^{-3}.
- g gravitational acceleration.
- α, β slope from horizontal of surface and base of thrust sheet.
- x_i Cartesian coordinate system, variously oriented.
- σ_{ij}, σ_{ij}^d, σ_{ij}', σ_m components of the full, deviatoric, supplementary, and mean stress tensors.
- τ basal shear stress.
- T variational stress.
- G stress deviator gradient.
- λ wavelength of irregularities at base of thrust.
- F longitudinal force along thrust.
- C compressive stress at boundary imposed by horizontal surface force.

H, L thickness and length of thrust sheet.
k ratio of C to $\rho g H \cos\alpha$.
τ_m maximum shear strength, equal to 2×10^7 Pa.
t time.
v mean horizontal velocity of thrust.
$B, \dot{B}$ erosion and time rate of erosion.
$e_{22}, \dot{e}_{22}$ vertical natural strain, vertical strain rate.
θ angle between least compressive principal stress σ_I and the x_i axis.

SI units have been used.

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