Thoughts on the tectonics of folded belts

A. W. Bally

SUMMARY: Balanced cross sections and their palinspastic reconstruction in structurally simple external zones of folded belts suggest that a limited amount of continental lithosphere was subducted (Ampferer or A-subduction). This process appears to be associated with preceding or synchronous basement remobilization. Therefore, the rigid or ductile nature and the timing of basement deformation remain as some of the most important orogenic problems. Decoupling and ductility contrasts within the lower continental crust and within the overlying sedimentary sequences are responsible for varying structural styles in mountain ranges.

Gravity gliding as an important factor for mountain building is examined in some detail. Soft sediment gravity tectonics on passive continental margins are dominated by listric normal growth faults. This style contrasts with observed styles of deformation in folded belts. Gravity tectonics induced by stretching of the underlying basement area is commonly observed during the rifting phase of passive continental margins and in episutural basins associated with orogenic systems.

Three opposing schools of thought are today proposing their images of mountain ranges:

The *fixists* (Beloussov 1962, 1975, 1977) visualize mountain building as the product of asthenospheric and related lithospheric diapirism. In their view, widespread thrusting and folding in mountain ranges is explained in terms of either gravity gliding or else gravity spreading (as defined by Price 1971, 1973).

Adherents of an *expanding earth* (Carey 1975, 1977) view mountain building as the fixists do, but essentially as an ensialic process; they differ from fixists because they see the origin of the oceans not by a process of oceanization but instead by accretion and spreading processes along mid-ocean ridges that record the vicissitudes of an expanding earth.

Finally, *plate tectonic* devotees see mountains as the product of subduction processes on converging plate boundaries; overthrust and folded belts are, in essence, of compressional origin and represent excess sediments and slices of the underlying crystalline crust that have been scraped off and decoupled from subducting lithospheric slabs. A large number of observations in mountain ranges can be well fitted into a plate tectonic frame of reference. However, it is only fair to state that despite the brilliant early intuitions of Ampferer (1906), Argand (1924), Staub (1928), and other alpine geologists, plate tectonics are not easily directly deduced from observations that are limited only to mountain ranges. The plate tectonics hypothesis remains anchored mainly in geophysical and marine geological observations.

Deep-sea sediments and possible remnants of oceanic floor (ophiolites) occupy, areally, only minor portions of folded belts. Consequently, a plate tectonic origin of mountain ranges is not all that obvious to an unbiased observer. Much of what we see today in mountain ranges suggests widespread mobilization and 'ductilization' of an earlier rigid sialic lithosphere. In other words, while the continental lithosphere of cratons remains rigid, the continental lithosphere of mountain ranges shows pervasive remobilization during orogenic processes, suggesting repeated lithospheric 'softening'.

In this paper I attempt to provide a perspective of the evidence for, and the relative roles of, normal faulting, thrust faulting, and folding in mountain ranges. Such a perspective may help in judging the realism of some of the geotectonic images mentioned previously. It must be realized, however, that the phenomena on which I propose to concentrate are but the near-surface expression of deeper seated igneous and metamorphic processes, which are much more difficult to unravel but so much more important for an understanding of orogeny.

Although this paper deals mainly with compressional tectonics and thrust faulting, most figures illustrate normal faulting. It is assumed that specialists in thrust faulting and nappe tectonics are familiar and have ready access to the relevant illustrations. However, the seismic examples of gravity-induced normal faulting included in this paper may suggest to folded belt specialists what to look for when they attempt to differentiate gravity from compressional tectonics.

Thrust and Nappe Tectonics. 1981. The Geological Society of London

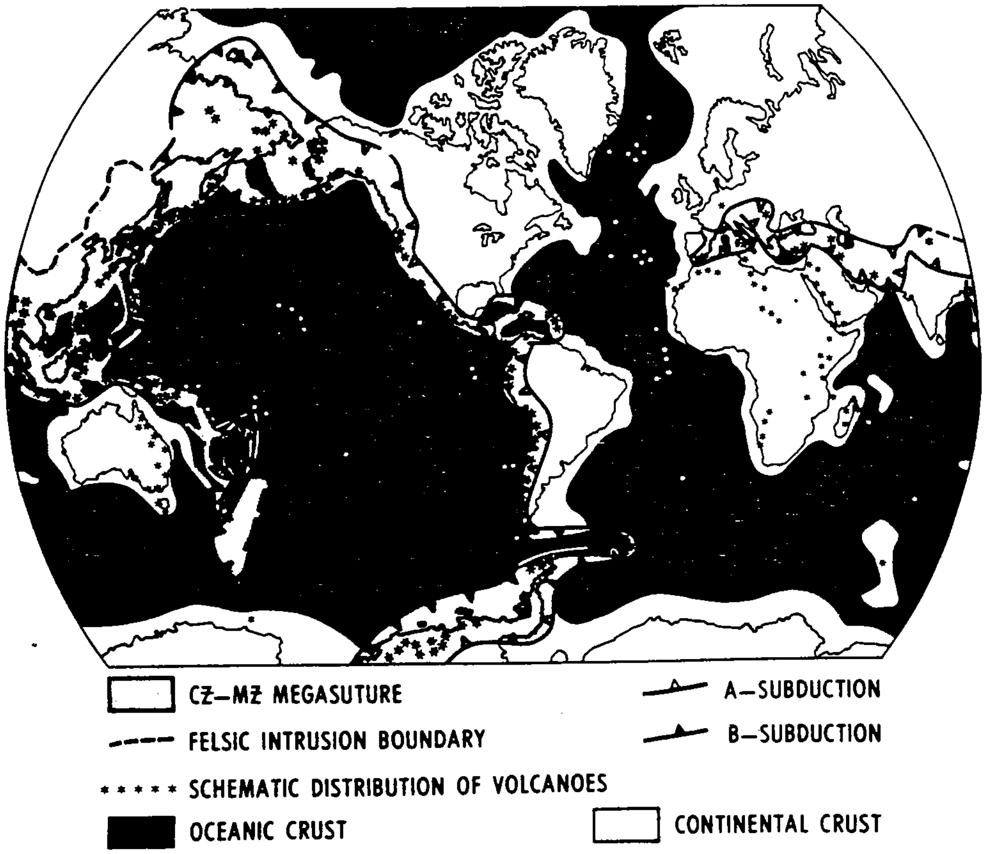


FIG. 1. Cenozoic–Mesozoic megasuture of the world and its boundaries. B-subduction zones face oceans; A-subduction boundaries face continental cratonic areas. They are characterized by widespread décollement folding and thrust faulting. In China the boundary is an ill-defined envelope around Mesozoic and Cenozoic granitic intrusives (after Bally & Snelson 1980, with permission of Canadian Society of Petroleum Geologists).

Without entering into detailed descriptions, Fig. 1 may serve to show a number of orogenic configurations. In earlier publications I proposed the term megasuture for foldbelts *and* sedimentary basins that are included in them (Bally 1975; Bally & Snelson 1980). For instance, the Mesozoic–Cenozoic megasuture includes all regions of intensive Mesozoic–Cenozoic mountain building and sedimentary basins involved in these processes. In plate tectonics jargon, the megasuture is the integrated product of all subduction-related processes which form the counterpart of the Mesozoic–Cenozoic ocean-spreading processes. The megasuture was introduced primarily to help in classifying sedimentary basins. The term also emphasizes the point that the bottoms of sedimentary basins within the megasutures are typically as deep as the adjacent mountains are high; and that the evolution of such basins has be be viewed as part of the total evolution of mobile belts (see Fig. 2).

Four types of megasuture boundaries are: B- (or Benioff) subduction boundaries where oceanic lithosphere is subducted; A- (or Ampferer) subduction boundaries where continental lithosphere is subducted; transform fault system boundaries; and—in China—a boundary which is an enevlope around felsic intrusives. This fourth boundary type is required because in China the continent facing boundary of the Mesozoic–Cenozoic megasuture is not associated with obvious external foldbelts involving former passive margin sequences and overlying foredeep sequences. Instead one can trace an ill-defined outline of Mesozoic and Tertiary intrusives which invade deeply into China and Mongolia.

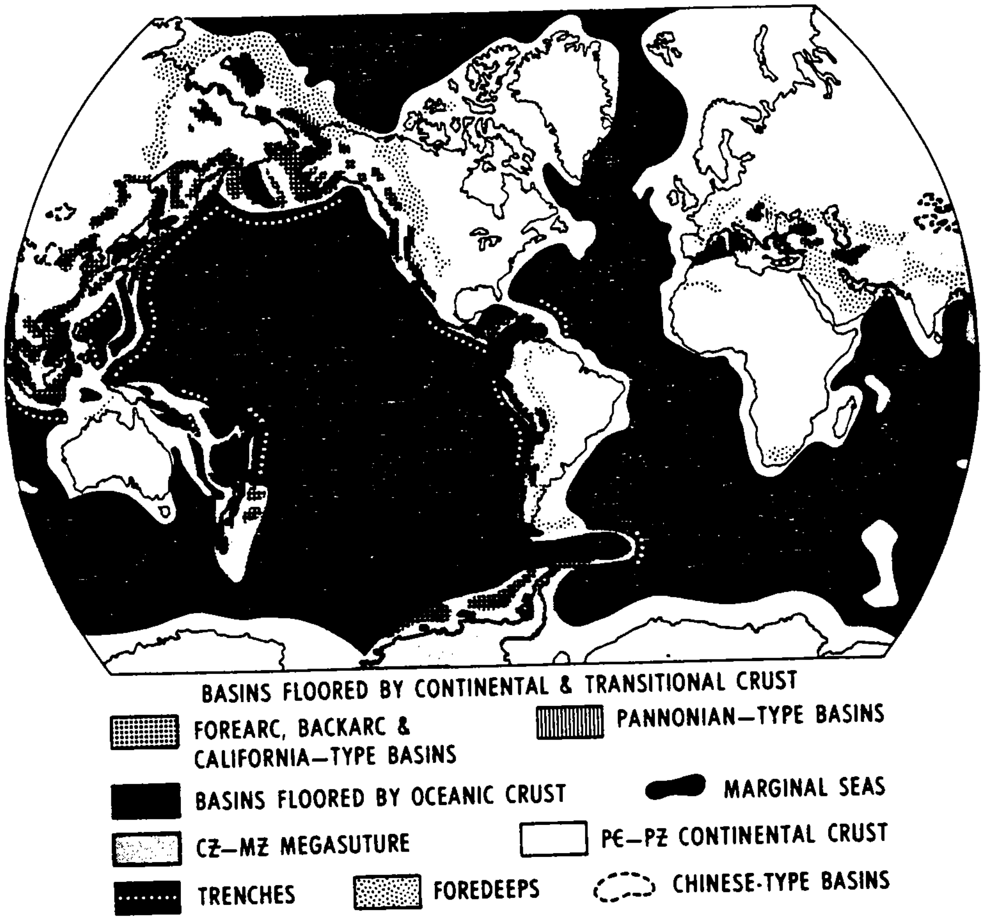


FIG. 2. Sedimentary basin families associated with the Cenozoic–Mesozoic megasuture (after Bally & Snelson 1980, with permission of Canadian Society of Petroleum Geologists).

With respect to the Mesozoic–Cenozoic megasuture, we differentiate four major 'orogenic' or megasuture types:

(1) *The SW Pacific type* contained between B-subduction and transform boundaries; a system of island arcs and marginal seas.

(2) *The NW Pacific type* contained between B-subduction and transform boundaries on the Pacific side and the felsic intrusion boundary of China. Within this orogenic type, marginal seas are opened and closed and continental fragments are captured (e.g. Indochina, South China platform, or the Lut Block of Iran).

(3) *The Cordilleran type* contained between a seaward B-subduction and/or transform boundary and a landward A-subduction boundary. The opening and closing of marginal seas appears to be less dominant in the history of the Cordilleras and it would appear that strike-slip rifting of continental fragments plays a major role.

(4) *The Alpine–Himalayan type* is contained within two A-subduction boundaries facing the Eurasian craton to the N and the African–Arabian and Indian continents to the S. This type is the end product of continental collisions.

Note that widespread overthrusting of former passive margin sediments and widespread regional basement remobilization and metamorphism is characteristically associated with A-subduction boundaries. B-subduction boundaries exhibit some imbricate thrusting, and deformation of oceanic sediments in the accretionary wedges of island arcs. Finally there is some thrusting associated with transform systems.

The case for compression and subduction in the external zones of foldbelts

External zones of foldbelts (the Externides of Kober 1928) are the accretionary wedges associated with Benioff subduction zones and more important, they are the folded belts that are associated with A-subduction zones.

The seismic record of Benioff zones favours extension and normal faulting in the peripheral bulge that is formed on the oceanward side of a deep-sea trench and compression in the shallow portions of the subduction zones and the associated accretionary wedge. Reflection seismic data on inner walls of deep-sea trenches suggest accretionary structures formed by 'offscraping' of oceanic sediments that overlie a gently dipping oceanic crust (Fig. 3). Scholl *et al.* (1977) noted that large volumes of pelagic sediments should be, but are not, exposed in Palaeozoic and Mesozoic Circum-Pacific mountain systems. These authors suggest that accretionary wedges may represent mostly deformed slope deposits and that much of the pelagic sediments were subducted with the oceanic crust.

Of course, one is tempted to construct balanced cross-sections (Dahlstrom 1969, 1970) in such a setting. However, the premises for the method are not easily fulfilled, because in most cases, the sediments that form the accretionary wedges are not dated by drilling, and further, because penetrative fabrics and the style of deformation indicates dominance of ductile flow and shear (for details, see von Huene *et al.* this volume). High ductility is suggested by outcrop studies of mélanges and the high pore pressures reported in some wells that are associated with subduction zones (Shouldice 1971). The situation is further complicated by an overprint of gravitational sliding and normal growth faults that show up on a number of reflection lines (see Colombia section, Fig. 3).

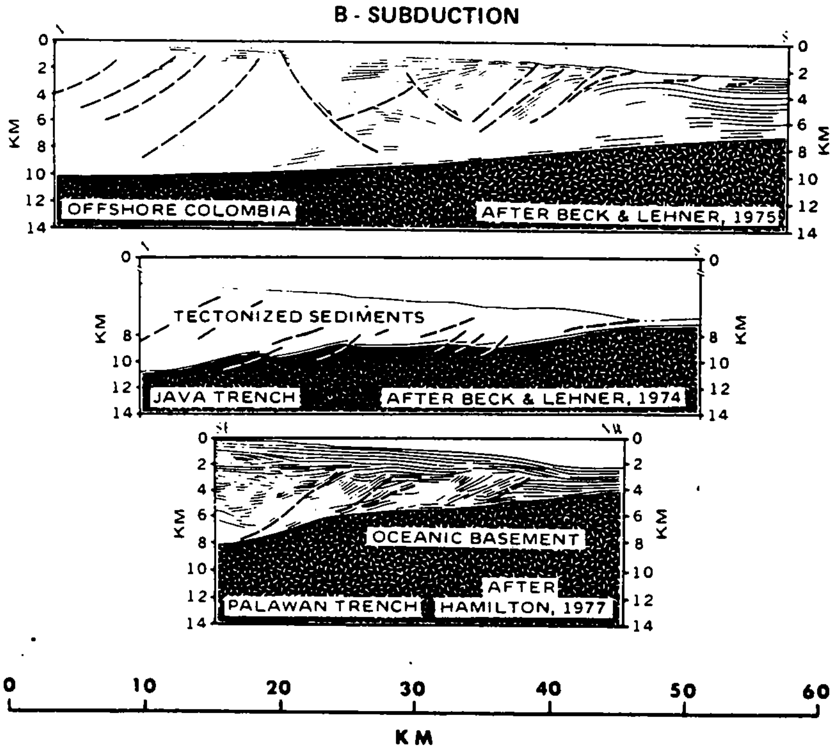


FIG. 3. Sketches of typical accretionary wedges associated with B-subduction zones. (a) Offshore Columbia (after Beck & Lehner 1975); (b) Java Trench (after Beck & Lehner 1974); (c) Palawan Trench (after Hamilton 1977); Drawings after Bally & Snelson (1980, with permission of Canadian Society of Petroleum Geologists).

Although one may rightly question the advisability of using balanced cross section techniques in Franciscan terranes, a recently published reconstruction of northern California by Suppe (1979) suggests some 175 km of shortening of Franciscan, a far cry from the shortening one would expect from offscraping of sediments during the subduction of an oceanic slab that was several thousands of kilometres wide.

One is forced to conclude that the current state of knowledge is inadequate to assess the amount of shortening associated with accretionary wedges of island arcs. The amount of underthrusting can only be derived from reconstructions that are based on magnetic stripe interpretations of the ocean floor. To understand structural deformation in accretionary wedges of island arcs and their ancient equivalents, it is increasingly more important to develope criteria that differentiate the effects of superficial gravitational sliding from the effects of compression and/or extension caused by the sinking oceanic lithospheric slab.

Shortening in folded belts associated with A-subduction boundaries has been studied for a number of years. In fact, the original concept of subduction in the Alps was based on such studies (Ampferer 1906; and later documented by more specific reconstructions: Spengler 1953–59; Trümpy (1969) has provided a detailed palinspastic reconstruction of the Glaronese Alps).

More accurate approximations to palinspastic reconstructions became possible with the aid of structural sections that were based on reflection seismic data. These data allow us to map the basement underlying the frontal folded belts in the Rocky Mountains (Bally *et al.* 1966; Price & Mountjoy 1970; Gordy *et al.* 1975; Royse *et al.* 1975) and in the Appalachians (Gwinn 1970; Roeder *et al.* 1978). A gentle mountainward dipping basement surface, observed on numerous reflection seismic sections constrains possible interpretations and leads to more rigid 'balancing' of cross sections as outlined by Dahlstrom (1969), Gwinn (1970), and Roeder *et al.* (1978).

Following Dahlstrom, balanced cross sections are only justified in a 'concentric regime', and the method needs modification if applied in a 'similar fold' regime. Bearing all cautions in mind, shortening in the Canadian Rockies exceeds 160 km and may reach 270 km (Bally *et al.* 1966; Price & Mountjoy 1970; also see Price this volume). For the Helvetic Nappes of the Alps and allowing for the uncertainties due to internal deformation, the accurate nature of tectonic units, and the estimation of eroded and buried parts, Trümpy (1969) estimates shortening in the order of 30–40 km. Spengler (1953–59) offers estimates typically in the order of 100–150 km for the northern Calcareous Alps.

Gwinn (1970) derives about 80 km for the Central Appalachians, a figure which is increased by Roeder *et al.* (1978) to about 140 km and which is to be further modified to in excess of 200 km in view of the new COCORP data in the area (Cook *et al.* 1979; Hatcher 1972).

For the Scandinavian Caledonides, which are at an erosional level that is particularly favourable for tracing the authochtonous foreland underlying higher thrust sheets, Gee (1975) postulates a total shortening in excess of 500 km. Similar amounts are postulated by Binns (1978) for the Caledonides of nothern Scandinavia. Note that Gee's figure is based mostly on credible inference from surface geology, without reflection seismic data. In the Caledonides like elsewhere, shortening estimates are critically dependent on resolving the penetrative strain recorded in allochthonous sequences, a point properly emphasized by Hossack (1978).

This author computed for the Jotun Nappe 65% vertical shortening and a transverse elongation of 160%. To restore this flattened nappe, its stratigraphic thickness needs to be multiplied by 2.15 to arrive at the original thickness. Even so, the displacement of the Jotun Nappe alone exceeds 290 km.

In other words, the amounts of shortening in external folded belts that are based on reasonably accurate geometric reconstructions and in some cases on reflection seismic data are typically in the range of 50–500 km. Even though in some folded belts of the world the amounts of shortening may be two or three times larger, the figures still are short of the several thousand kilometres that may be deduced from plate motions based on magnetic stripe reconstructions or else on palaeomagnetic data. This suggests either that plate tectonic reconstructions are basically incorrect or else that the subduction mechanism is most efficient in destroying much of the evidence which could be used for a quantitative kinematic check of reconstructions that are based on palaeomagnetics. Because of the convincing nature of the palaeomagnetic data, I favour the second alternative.

For an understanding of mountain building and the genesis of continental crust, it is desirable to examine the fate of the basement which was originally underlying the excess sediment now piled up in thrust sheets, nappes and folds. All authors who have actually carried out quantitative reconstructions in the external zones of folded belts agree that the crust underlying the excess sediments remained in the subsurface: that is, nowhere has a corresponding tectonically denuded surface been exposed, nor is there evidence for subaerial exposure of such a surface that was followed by subsequent burial. Consequently, gravitational gliding will have to be ruled out as a major factor in mountain building.

The concept of gravitational spreading as developed by Price & Mountjoy (1970), Price (1971, 1973), and Elliott (1976*a*, *b*) suggests that gravitational forces dominate the emplacement of thrust sheets and, following Elliott, that significant surface slopes are required to form thrusted and folded mountain belts. The linkage of folding in the external zones with the emplacement of metamorphic folds of the Pennine Nappe type is not clear and adequately documented because in a number of cases, such metamorphism precedes the deformation in the external foldbelts.

To conclude: the basement originally underlying excess sediments of external foldbelts remained at depth either to form a mountain 'root' leading to formation of a thickened lithosphere or else it was engulfed in the overall lithospheric subduction process which led to the formation of mountains. In all cases the apparent disappearance in depth of continental lithosphere (A-subduction) is limited to hundreds of kilometres and contrasts with B-subduction involving disposal of thousands of kilometres of oceanic crust. There is little doubt that surface slopes are increased during both A- and B-subduction processes. Such slopes may facilitate gravity spreading. However, the proof for gravity spreading rests on demonstrating that diapir-like metamorphic structures in the internal zones of folded belts formed at the same time as the foreland folds and thrusts which record substantial shortening, and also on a demonstration that extensional tectonics in more brittle overlying sequences are synchronous with the shortening in the foreland. To my knowledge, such conclusive proofs have yet to be published.

This paper is not concerned with the mechanics of the A-subduction process. Conceptual and mechanical difficulties of subduction of substantial portions of continental lithosphere are in part overcome by the simple observation that in all cases we deal with a presumably attenuated continental crust of the lithosphere of a former passive margin. Nevertheless, until recently, there was a great deal of reluctance by plate tectonic experts to accept A-subduction. It has been argued that the high buoyancy of continental crust would prevent significant subduction of the continental lithosphere. Molnar & Gray (1979) calculate that significant fractions of the continental crust may be subducted if these could be detached from their upper part. Furthermore, these authors suggest that the gravitational force acting on sinking oceanic lithosphere may pull continental lithosphere into the asthenosphere. Such a pull is counteracted by the buoyancy of the light continental crust. Under these circumstances and using varying assumptions, typical values between a few kilometres and up to 330 km in length of subducted continental crust appear to be reasonable. These calculations depend on the thicknesses of lower continental crust that may be detached during the subduction process. If somewhat more extreme assumptions are used, much greater lengths of continental lithosphere may be subducted.

The model of Bird *et al.* (1975) and Bird (1978) provide a thermal and mechanical scenario which leads to the 'delamination' of sub-crustal lithosphere by insertion or wedging of less viscous asthenosphere between an upper crustal layer and the underlying denser subcrustal lithosphere.

While there is little to add to these interesting models, it is of some comfort to know that subduction of continental lithosphere (A-subduction) which for some time was repulsive to theoretical plate tectonicians has now been accepted in principle. Thus, as geologists, we may now continue to gather observations that may bracket the actual amount of shortening observed in mountain ranges.

The phenomenology of gravity tectonics of sedimentary sequences

Although all preceding considerations seriously limit the role and significance of gravity gliding processes for mountain building, it still may be useful to develop criteria for recognition of gravity tectonics by looking at some obvious examples.

Reviews of gravity gliding have been offered by de Sitter (1954) and North (1964). Various aspects of gravity tectonics are dealt with in a book dedicated to van Bemmelen (de Jong &

Scholten 1973). While many of these authors spend a great deal of effort to convince the readers of the correctness of their mountain building images, very little is offered by way of description of large regions that are unambiguously dominated by gravity tectonics. Two examples of such areas are the Niger Delta and the northern Gulf of Mexico.

The Niger Delta, as described by a number of authors (e.g. Delteil *et al.* 1976; Weber & Daukoru 1976; Lehner & de Ruiter 1977), is prograding on a foundation of high pore pressure shales that overlies an oceanic crust. It is characterized by extensive growth fault systems and shale diapirism. Particularly the toe of the Delta is characterized by imbrications that are similar to features often observed in folded belts (Lehner & de Ruiter 1977).

Like the Niger Delta, the Gulf Coast Tertiary is also prograding on a high pore pressure shale substratum that in turn appears to be underlain by Mesozoic carbonates and a basal salt-bearing sequence which provides an additional unstable base. Where penetrated by the drill, the basement appears to be continental and an extension of the Palaeozoic mountain system of the Ouachitas and the Appalachians. However, moving towards the Gulf, the crust changes from continental to transitional and/or oceanic (Fig. 4). The structural evolution of the Gulf of Mexico is far from being unraveled, but a few characteristic details shown in reflection seismic lines may offer useful reminders for geologists interested in gravity tectonics in folded belts and may suggest what to look for in support of gravity-gliding concepts.

As suggested on Fig. 4, the edge of the salt mass appears an allochthonous glacier-like tongue overlying very young Tertiary sediments (Watkins *et al.* 1978; Humphris 1978). Widespread pre-Cretaceous listric normal growth faults are restricted to the sedimentary sequence and flatten out at the base of the salt (Figs 5, 6, & 7). Farther up in the sequence and towards the Gulf of Mexico, extensive listric normal growth faults dissect Tertiary clastics and flatten within the high pore pressure shale section at depth. Diapiric salt movement and questionable shale diapirs are intimately associated with these growth fault systems (Figs

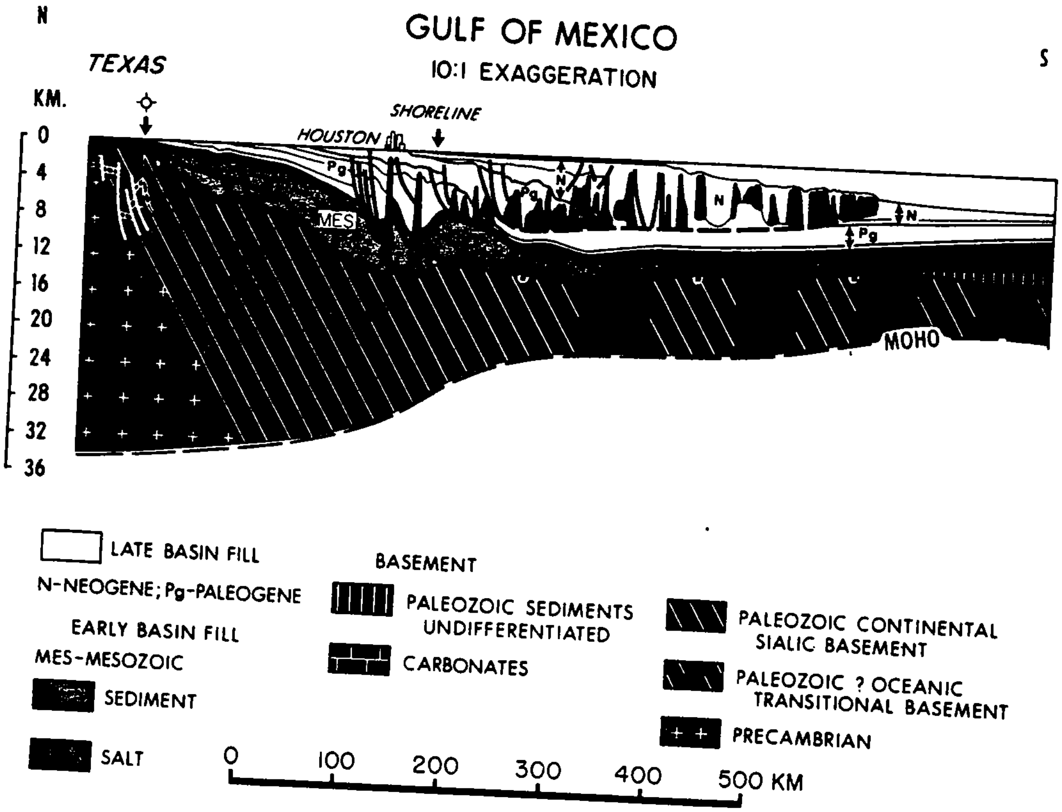


FIG. 4. Schematic section across the Gulf of Mexico.

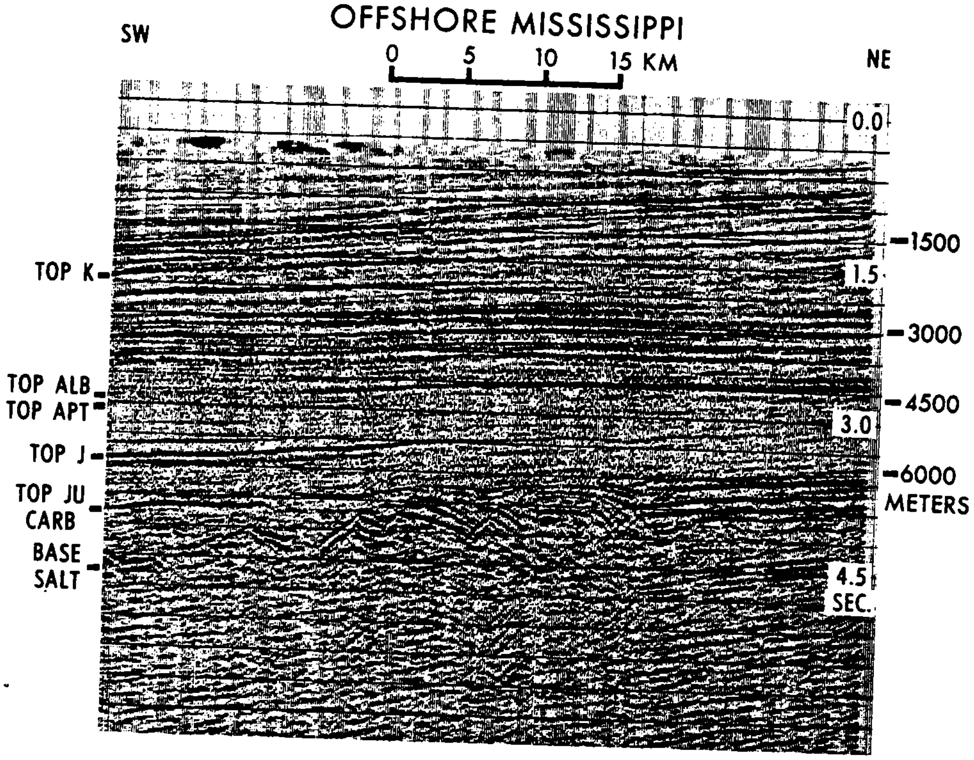


FIG. 5. Offshore Mississippi, reflection seismic section. Note listric normal faults separating salt rollers involving Jurassic carbonates.

8 & 9). The interaction of sedimentation and growth faulting in this region has been summarized by Curtis (1970) and Curtis & Picou (1978), and its relevance for the genesis of hydrocarbon deposits is well illustrated by Curtis (1979). Detailed studies of listric normal faulting and sedimentation are also known from western Ireland (Rider 1978) and Spitzbergen (Edwards 1976).

Gravity tectonics of the type indicated are typical in areas underlain by soft sediments, unstable high pore pressure shales and salt that are associated with high rates of subsidence. Listric normal growth fault systems can be observed in depths in excess of 20 000 ft (6000 m).

An example which shows listric normal faulting that is more or less synchronous with compressional folding is seen on Mexico's eastern offshore (Buffler *et al.* 1979; Watkins *et al.* 1976; Fig. 10). Although more data and calibration by drilling are needed, it appears possible that the amount of shortening represented in the folds of the Mexican ridges may correspond roughly to the amount of stretching by listric normal faulting underlying the Mexican shelf. On the other hand, if the amount of shortening of the linear folds substantially exceeds the postulated stretching, we may interpret the data as a much younger equivalent (i.e. Pleistocene) of the Laramide Sierra Madre Oriental folds, onshore to the W.

Friends of gravity gliding will be quick to point out that—contrary to the Gulf of Mexico example—much of the deformation occurring in folded belts affects already lithified sediments and in several cases the underlying basement. Consequently, it is also of some importance to characterize clearcut gravity tectonics in 'hard' rocks.

Some very well documented examples of local superficial gravitational gliding have been given by Pierce (1957, 1963, 1973) for the Heart Mountain area of Wyoming and by Reeves (1946) for the Bearpaw Mountains of Montana.

Normal faults responding to extension of the upper crust are most common in the western Cordillera of the USA and the southernmost segment of the Canadian Rocky Mountains.

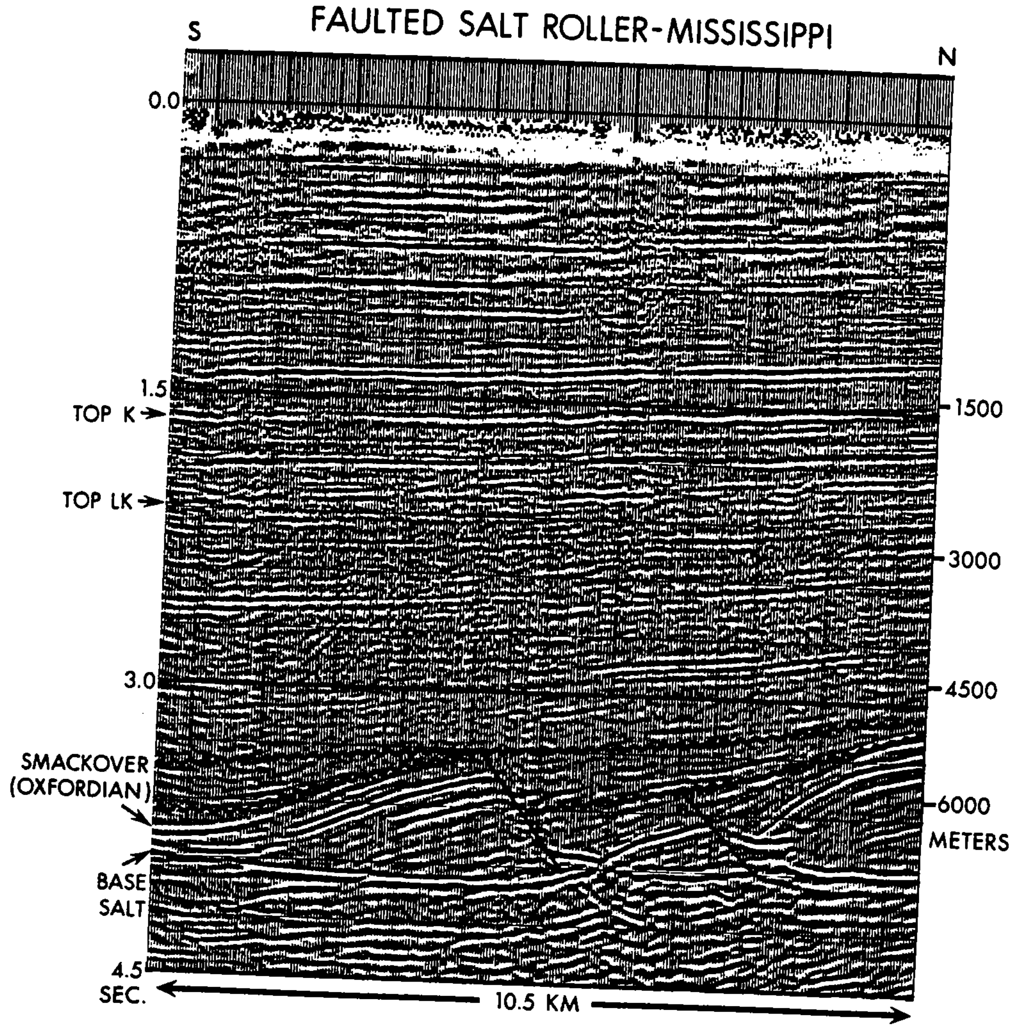


FIG. 6. Gulf Coast in Mississippi, reflection seismic section. Salt rollers, bounded by normal faults which flatten and merge with base of the salt.

Here again, reflection seismic data indicate the listric nature of the faults. The evidence for this is the presence of continuous reflections underlying obvious normal faults that are post-thrusting in age and can be mapped on the surface. The presence of such reflection data does not permit the direct straight-line projection into the subsurface of the large fault offsets seen at the surface; instead the normal faults have to flatten quickly at depth (see Fig. 11; Bally *et al.* 1966; McDonald 1976).

Another expression of the listric nature of these normal faults is the widespread rotation into the fault plane of beds that were deposited while the fault was active (Fig. 12). In the western Cordillera the evidence for the listricity of normal faults ranges from superficial low-angle normal faults that simulate the base of a major landslide system to intermediate depth fault systems and exhumed, formerly deep fault systems that separate the more ductile deformation realms of metamorphic core complexes from the brittle overlying sediment cover (Davis & Coney 1979; Davis in press; Effimoff & Pinezich, 1980).

The normal fault systems in the western Cordillera are part of a very complex megashear system that—most unfortunately for 'gravity gliders'—clearly postdate the major overthrusting events of the western

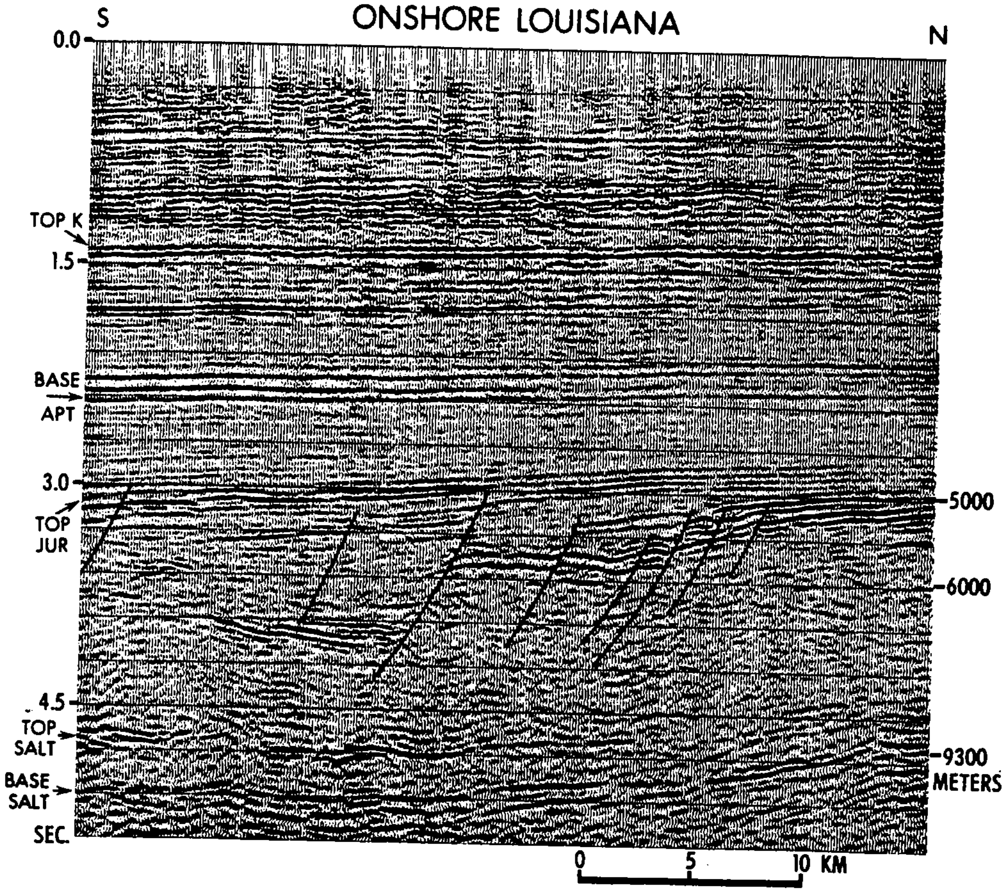


FIG. 7. Onshore Louisiana, reflection seismic section. Note pre-Cretaceous listric normal faults.

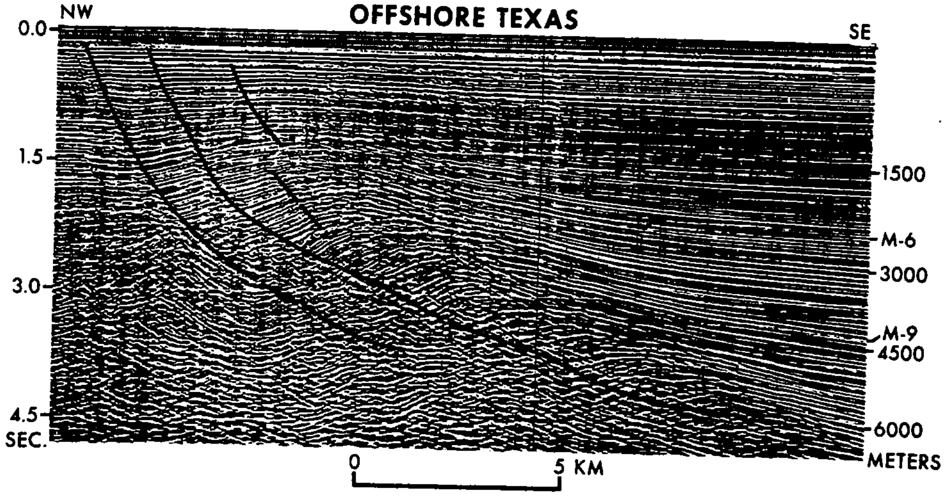


FIG. 8. Offshore Texas, reflection seismic section. Note extensive listric normal growth faults in Miocene section. M-6, M-9 are Miocene marker beds.

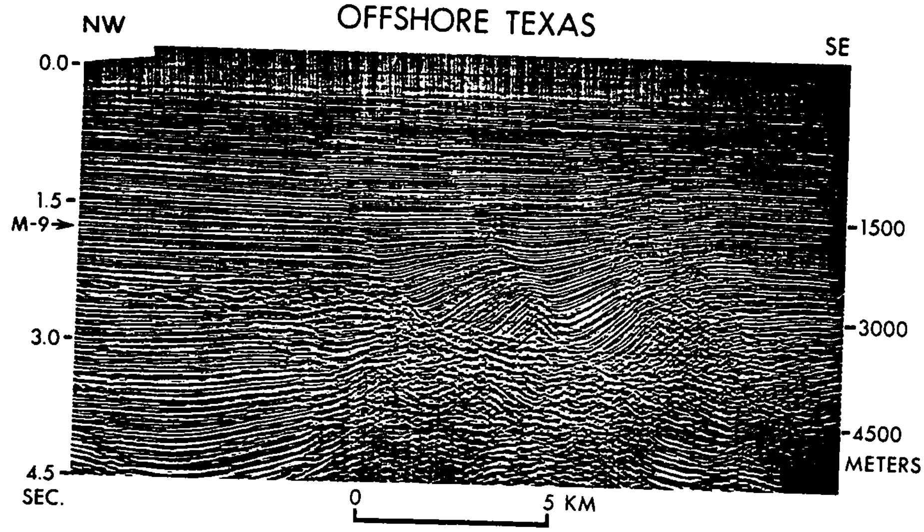


FIG. 9. Offshore Texas, reflection seismic profile showing the interaction of growth faults and diapiric structures. M-9 is a Miocene marker bed.

Cordillera. Note, however, that in a somewhat similar setting, extensional gravity tectonics and normal faulting are also common in the Vienna and the Pannonian Basins (Prey 1974). There, however, the normal faulting roughly occurs during the same time brackets as the last phase of overthrusting in the adjacent Carpathians. However, traditional gravity gliding is precluded by the simple fact that inner portions of the Carpathians are typically subsiding during the Tertiary, instead of furnishing an elevation from which nappes could glide towards the adjacent foredeep.

We may conclude that if we are to explain extensive thrusting and the formation of nappes by gravity gliding, we ought to search for extensive listric normal faulting that is synchronous and directly associated with the overthrust phenomena we observe. The amount of extension of these faults should be comparable to the amount of compressional shortening in the associated foldbelt. So far, I have failed to find in any mountain ranges evidence that would so link the emplacement of overthrust sheets with commensurate extensional fault systems in the inner portions of orogenic belts.

Thrust faulting involving the crystalline basement

So far, the discussion has been concerned with the evidence for the relative role of gravity gliding versus shortening by subduction-related compression. Support has been obtained by studying surface data and reflection seismic data that illustrate décollement tectonics of sedimentary sequences. A few words concerning the involvement of continental crystalline basement in orogenic processes is now in order.

One group of examples of orogenic basement involvement includes cases where slabs of varying thicknesses of, more or less, rigid crystalline basement rocks form thrust sheets. Examples are the eastern Alpine thrust sheets, the Main Central Thrust of the Himalayas, the Blue Ridge of the Appalachians, the Caledonides of Norway, and many others. In this first group, the basement has not been pervasively remobilized and contrasts with a second group of examples where the crystalline basement was extensively remobilized as in the Penninic Nappes of the Alps or the Shuswap complex of the Canadian Cordillera. Both groups can be viewed as end members for transitions showing varying degrees of basement mobility have been mapped across many folded belts. Instead of reviewing in detail the structural deformation of crystalline basement in folded belts, only the following points will be re-emphasized:

The involvement of crystalline continental basement in folded belts is characteristic and

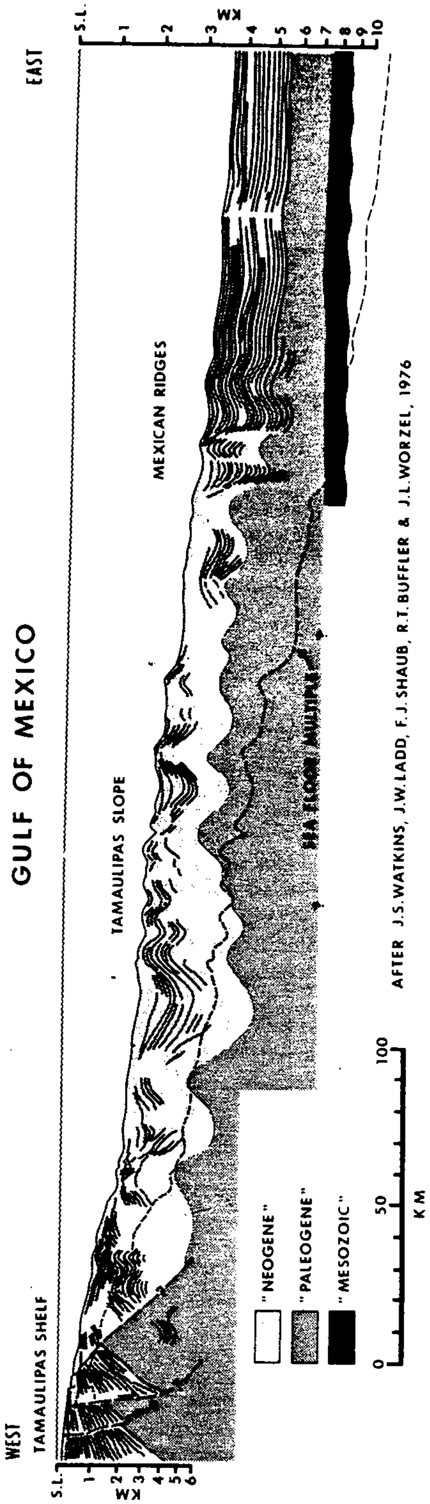


FIG. 10. Sketch of seismic line across Mexican Ridges (after Watkins *et al.* 1978). Note growth faults on left side of section and compressional décollement folds on the right side of the section.

constitutes the main reason why folded belts are often regarded as the product of ensialic orogenies. Some authors see a contradiction between a purely ensialic orogeny and orogenies that are related to plate tectonic processes. It is true that a number of mobile belts involve oceanic crust or ophiolites in suture zones. These permit us to postulate oceans of unknown width in plate tectonic reconstructions. However, in other cases, the evidence for such oceans is not so obvious, and at least in one case, folding and associated basement remobilization is entirely ensialic (Amadeus Basin of central Australia: Wells *et al.* 1970). Admittedly, the last example is unusual, because it affects a relatively small inverted basin located in the foreland of the Tasmanides. Nevertheless, the Amadeus Basin appears to offer an unambiguous case of ensialic deformation with décollement folding accompanied by the formation of small crystalline nappes.

Thrust faulting involving basement clearly indicates the existence of decoupling levels within the continental lithosphere. These may be mostly in the lower crust or may be in the upper mantle. Recent geophysical work in the Alps and Appennines (Angenheister *et al* 1972; Giese *et al.* 1973, 1978; Mueller *et al.* 1976; Mueller 1977) gives evidence for widespread low-velocity layers within the crust that extend well into the foreland. Thus the case for a layered crust with potential intracrustal decoupling levels is reinforced. The rheologic characteristics of these low-velocity layers are as yet poorly defined and consequently, the genesis of crustal low-velocity layers remains a matter for speculation. Some of the geological consequences of this problem are discussed by Hsü (1979).

There is also evidence for widespread crustal decoupling in the foreland of folded belts. A recent reflection line across the Wind River Mountains of Wyoming (Smithson *et al.* 1978, 1979; Brewer *et al.* in press) indicates that this basement uplift is underlain by a thrust fault which can be followed to a depth of about 35 km (see Fig. 13). If—as I believe—this line is characteristic for foreland block faulting, then it would follow that a major Laramide decoupling zone occurring at or near the base of the continental crust was underlying the Cordilleran foreland or the central and southern Rockies of the USA.

In a more speculative vein, attention should be called to Ziegler's (1978) observation that the well-known inversions observed in northwestern Europe (e.g. Wealden Anticlinorium,

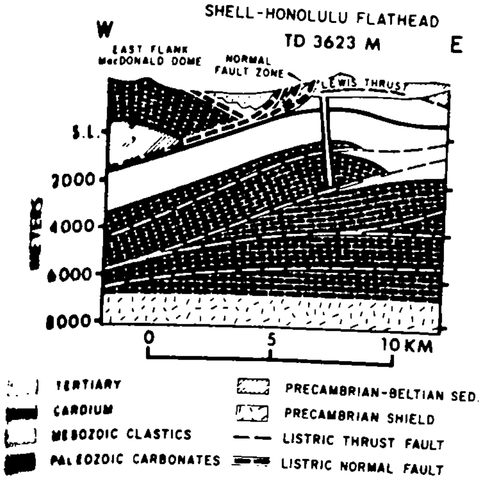
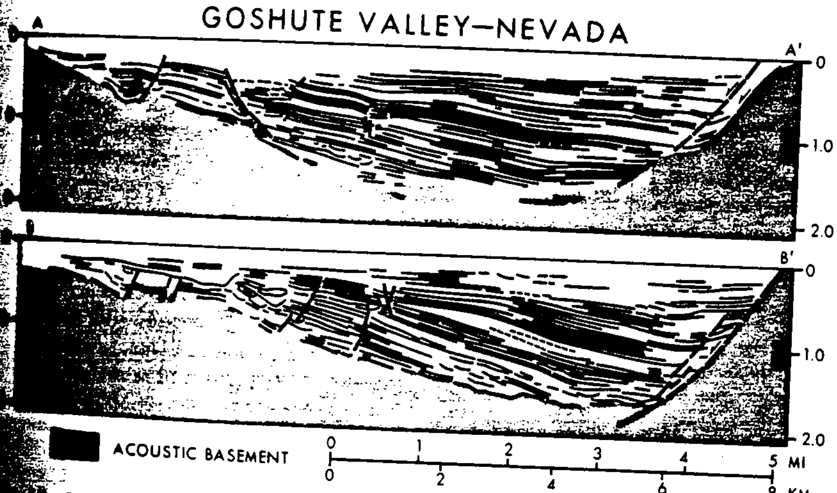


Fig. 11. Post-orogenic normal faults in Flathead area of British Columbia (from Bally *et al.* 1966, with permission of Canadian Society of Petroleum Geologists).

Netherlands Basin, Lower Saxony Basin, the Polish Anticlinorium, etc.) occur dur- the Meso-alpine deformation. This may mean that with the incipient alpine colli- stresses are transmitted over more than km across a rigid basement plate that a deeper, less competent decoupling Such a level may well be located within wer crust or the upper mantle. The con- further supported by structural observa- (horizontal stylolites, joints, minor reverse and strike-slip faults) in the sediment cover of the intervening platforms (Wagner 1974; de Charpal *et al.* 1974; Wunderlich 1974; for an updated review, see also Letouzey & Tremolières 1980).

There is a great deal of similarity between the foreland tectonics described in the preceding paragraphs and the continental collision processes that were described by Molnar & Tapponier (1978) and Tapponier & Molnar (1976) for Central Asia. There the late Palaeogene–Neogene collision of India with Eurasia caused extensive strike-slip faulting, thrust faulting, and normal faulting in Central Asia, which according to Molnar & Tapponier led to much of China being squeezed in an eastward direction. It should be noted that the collision also led to the emplacement of basement thrusts in the Himalayas, and it appears likely to me that some thrust faults associated with the Tien Shan and the Nan Shan uplifts involve thick segments of the continental crust in a manner comparable to the Rocky Mountain foreland.

It is concluded that decoupling within the lower crust and perhaps in the upper mantle is essential to explain thrust faulting of continental basement slabs. More geophysical information is needed to map such decoupling levels. These may well coincide with low-velocity layers that have been determined in a number of crustal studies. We are accustomed to explaining varying styles of deformation in folded and thrust-faulted sedimentary sequences as a function of ductility contrasts that are inherent



12. Sketch of seismic line across Goshute Valley, NE Nevada (after Bally & Snelson with permission of Canadian Society of Petroleum Geologists).

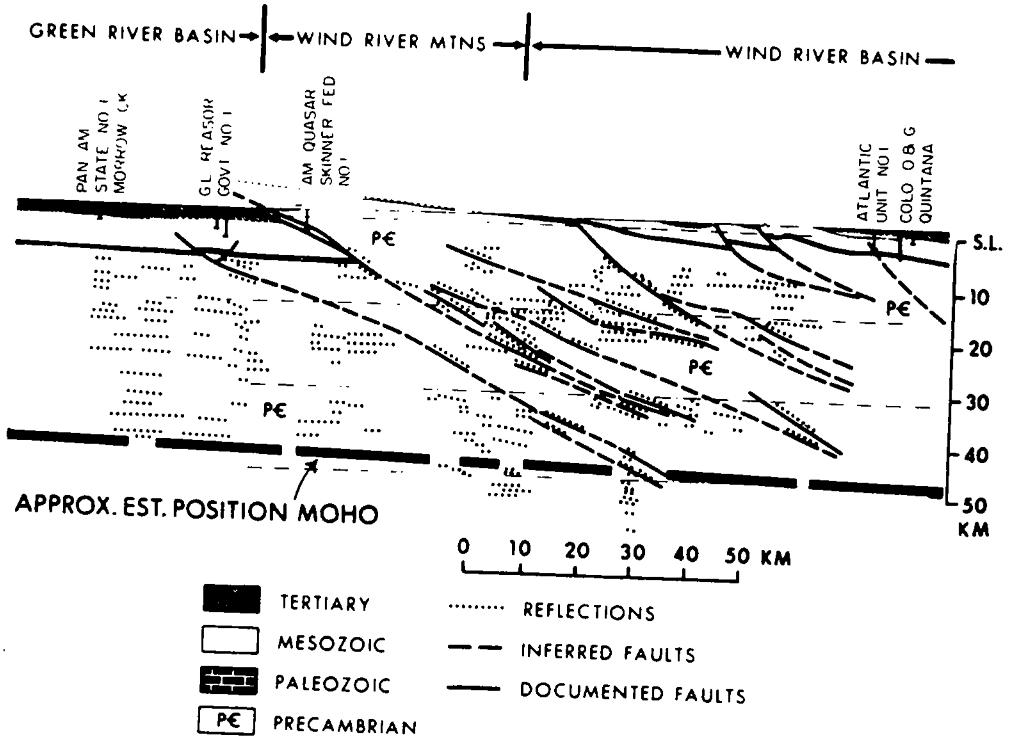


FIG. 13. Interpretation of Wind River Mountains, Wyoming, based on COCORP line (Smithson *et al.* 1978).

in the stratigraphic layering of the deformed sequences. By analogy, the style of deformation of basement slabs is dependent on ductility contrasts within the crystalline basement.

A number of plate tectonic models imply crustal decoupling (Armstrong & Dick 1974; Oxburgh's flake tectonics 1972; Bird's delamination 1978; Molnar & Gray 1979). A significant increase in understanding requires that future work concentrates on getting more detailed geological and geophysical documentation of deep crustal decoupling.

Turning to basement mobilization in the inner folded belts, it may be said that the causes for widespread regional remobilization and metamorphism of the basement remain obscure and may be due to burial and loading by higher thrust sheets or possibly due to thermal uplifts.

Overthrusting and strike-slip faulting

In recent years a number of authors have discussed the nature of thrust faulting associated with strike-slip faulting (Lowell 1972; Wilcox *et al.* 1973; Harding 1973, 1974, 1976; Sylvester & Smith 1976; Harding & Lowell 1979). These authors document and systematize the phenomenology of wrench faulting with surface, subsurface examples and clay model studies. An important problem relates to the scale and importance of thrust faulting, associated with strike-slip faulting. Clearly *en échelon* faults and thrust faults occur in a wrench fault regime. The upthrust interpretation proposed by various authors needs more verification by reflection seismic data. There remains, however, the question of how much shortening of sediments can be taken up by strike-slip faulting.

A schematic cross-section and reconstruction made by my colleague, R. E. Farmer, illustrates the problem (Fig. 14). The Taiwan foldbelt has a structural style that appears to be similar to the Rocky Mountain Foothills of Canada and the outer Carpathians of Rumania. Although no seismic data concerning the underlying basement are published, one may assume gentle eastward dip by the reconstruction of the stratigraphic wedge that

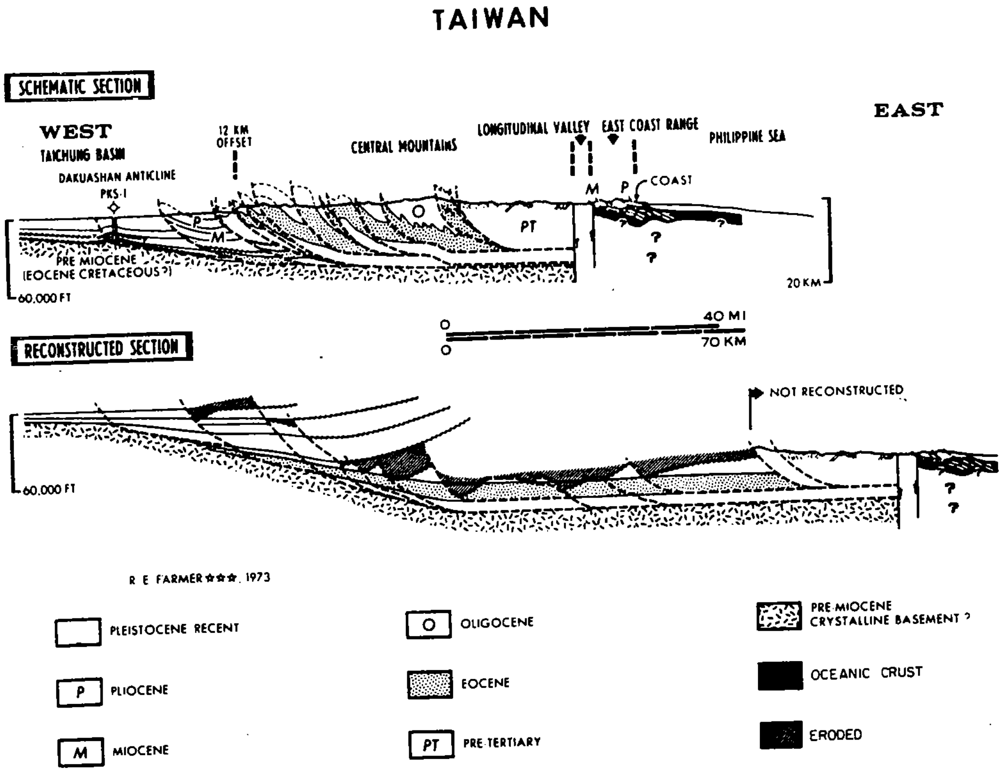


FIG. 14. Schematic cross section and reconstruction across Taiwan by R. E. Farmer (Shell Oil Company). *** Indicates quality.

is involved in the deformation. As a 'denuded' basement does not outcrop in the adjacent mountains, conventional gravity gliding may be precluded.

Accepting the admittedly shaky premises of the cross section, one has to conclude that a strip more than 50 km wide of pre-Tertiary basement apparently was subducted to form a deep 'root' under Taiwan. The crustal character of that basement is unknown. The subduction of the basement could have preceded the strike-slip displacement along the fault of the Longitudinal Valley or else the subduction process occurred during the deformation of that fault and during the folding in western Taiwan. Phases of subduction alternating with strike-slip faulting over short geological time spans can also be imagined.

The conclusion is that a substantial room problem may occur in palinspastic reconstructions of balanced cross sections across foldbelts that appear to be related to or later modified by strike-slip fault systems. Obviously, more reflection seismic data are needed to gain a better feeling for the dimension of the problem. Plate tectonic reconstructions based on palaeomagnetic data frequently suggest the location of orogenic systems in an overall strike-slip/shear context and therefore it becomes increasingly more important to differentiate and determine the scale of thrusting related to strike-slip faulting, A-subduction and B-subduction.

Folded belts, basement mobilization and plate tectonic reconstructions

A corollary of many of the preceding comments and the megasuture concept is that the crystalline basement of folded belts did not behave as part of a rigid lithosphere during orogenic processes. Consequently, it is of some importance to separate relatively more rigid lithospheric realms from basement that has been remobilized and subjected to regional metamorphism and from basement fragments that were overthrust or else were rifted by complex strike-slip movements.

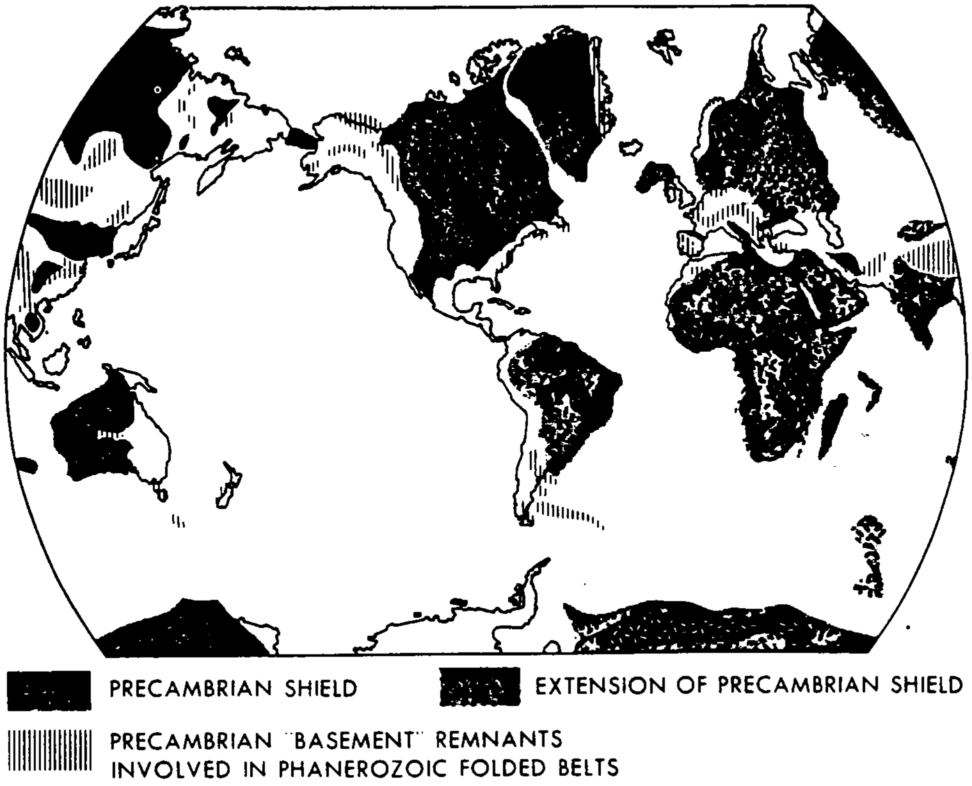


FIG. 15. Preserved Precambrian plate remnants. Precambrian 'basement' remnants that have been remobilized within Phanerozoic foldbelts do not qualify as 'rigid' microplates.

Superficial décollement folds and thrust sheets of Externides (Kober 1928) are probably in most cases underlain by a gently dipping basement ramp which represents unambiguous rigid lithosphere. The inner remobilized and metamorphosed portions of folded belts and small interior basement blocks captured during the orogenic process obviously represent either 'ductilized' lithosphere or else stray crustal fragments.

Plate tectonic reconstructions that are based on palaeomagnetic or stratigraphic points of control which are located within folded belts obviously have to be contrasted and related to lithospheric 'cratonic' segments that are now adjacent to these foldbelts. Therefore, there is an urgent need for reasonably accurate structural reconstructions of folded belts. At the same time, it is desirable to map the outlines of pre-Mesozoic and Precambrian continental lithospheric remnants to provide some of the building blocks for plate tectonic reconstructions. Outlines of the different types of building blocks are shown on Fig. 15 for the beginning of the Palaeozoic and on Fig. 16 for the beginning of the Mesozoic. The extensive occurrence of Precambrian and Palaeozoic basement remnants in later folded belts again emphasizes the 'ensialic' aspects of mountain building.

Conclusions

Normal faulting in folded belts and their foreland may be: —listric normal faulting involving the basement and related to the passive margin phase preceding orogenic deformation (for analogue, see de Charpal *et al.* 1978); —surficial soft-sediment listric normal faulting related to the drifting phase preceding orogenic deformation; —listric normal faulting related to the genesis of synorogenic accretionary wedges of B-subduction zones; —synorogenic normal faulting in the foreland related to the foreland bulge associated with subduction zones (Buchanan & Johnson 1968; Hopkins 1968; Laubscher 1978); —syn or

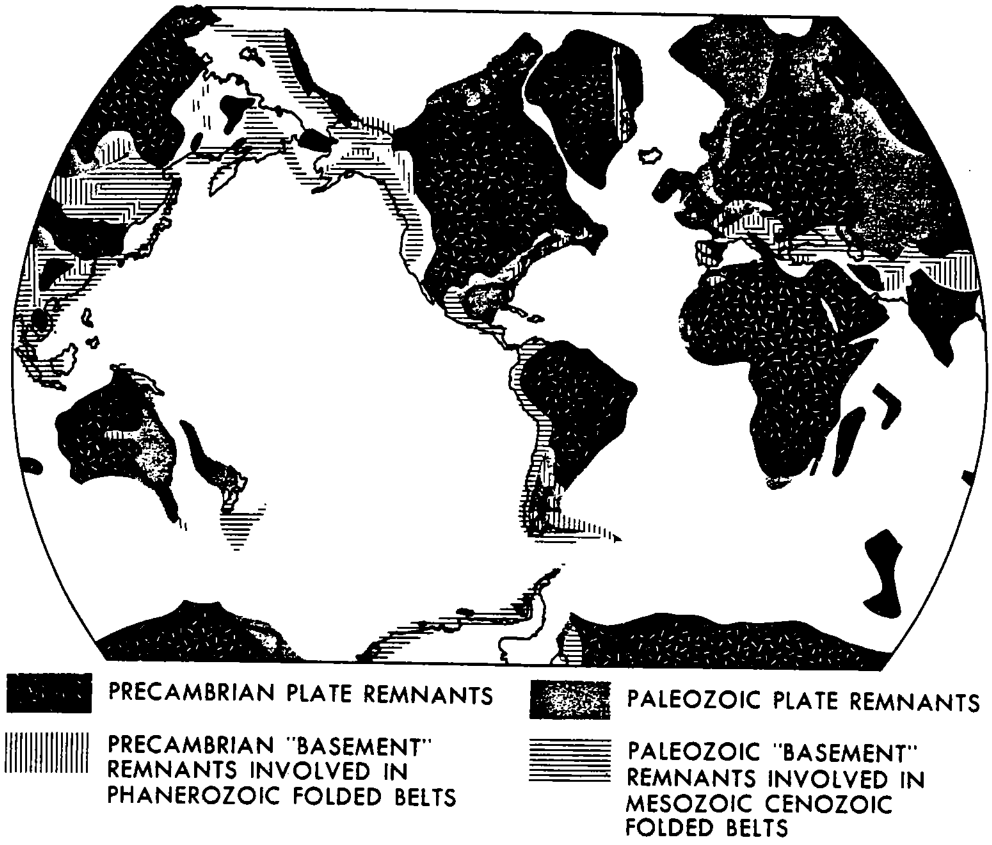


FIG. 16. Preserved Precambrian and Palaeozoic plate remnants. Palaeozoic and Precambrian outcrops that have been remobilized within Phanerozoic foldbelts do not qualify as preserved "rigid" microplates.

postorogenic listric normal faulting associated with stretching and shearing of the orogenic system (e.g., Great Basin, Vienna and Pannonian Basins).

Thrust faulting in the folded belts may be: —minor preorogenic thrust faulting at the toe of deltaic systems of passive margins (for analogue, see Lehner & de Ruiter 1977); —synorogenic listric thrust faulting involving the continental basement within the mountain range (e.g. Himalayas and eastern Alpine Nappe) and in the foreland of mountain ranges (e.g. Wind River Mountains); —synorogenic thrust faulting or sedimentary sequences related to A-subduction processes (e.g. Canadian Rocky Mountains, Appalachians, and Externides of the Alpine system); —synorogenic thrust faulting of sedimentary sequences related to B-subduction processes (see Fig. 3); —synorogenic thrust faulting related to strike-slip faults (Fig. 14).

All information on folded belts permits and supports their plate tectonic origin at subduction or else at transform plate boundaries. However, much of what we actually observe in mountain ranges involves décollement of and within sedimentary sequences, as well as significant decoupling within the deeper continental crust. Normal faulting and thrust faulting due to gravity gliding is not important for mountain building. This and the absence of wide-spread tectonically denuded basement argues strongly against a major role for gravity gliding.

Future geological and geophysical work should aim at defining and mapping such decoupling levels, particularly those that occur within the crust. Seismic crustal studies are a particularly promising aid in mapping intracrustal decoupling levels.

Acknowledgments. I would like to thank my colleagues R. E. Farmer for providing the geological cross section of Taiwan, R. L. Nicholas for providing the Gulf Coast reflection lines, and K. Arbenz and S. Snelson for reviewing the paper and for their helpful suggestions. I also thank Shell Oil Company for permission to publish the seismic sections and the paper.

References

Ampferer, O. 1906. Über das Bewegungbild von Faltengebirgen, Austria. *Jahrb. geol. Bundesanst,* **56,** 539–622.

Angenheister, G., Bögel, H., Gebrande, H., Giese, P., Schmidt-Thomé, R. & Zeil, W. 1972. Recent investigations of surficial and deeper crustal structure of the eastern and southern Alps. *Geol. Rdsch.* **61,** 349–95.

Argand, E. 1924. La tectonique de l'Asie, *Compt. Rend III^e. Congr. Int. geol. Liège,* Imprimerie Vaillant-Carmanne (transl. & ed. by Carozzi, A. V.). Hafner Press, New York, 218 pp.

Armstrong, R. L. & Dick, H. J. B. 1974. A model for the development of thin overthrust sheets of crystalline rock. *Geology,* **2,** 35–40.

Bally, A. W. 1975. A geodynamic scenario for hydrocarbon occurrences. *Proc. Ninth World Petrol. Congr. Tokyo,* **2,** Applied Science, Essex, 33–44.

——, Gordy, P. L. & Stewart, G. A. 1966. Structure, seismic data, and orogenic evolution of Southern Canadian Rocky Mountains. *Bull. Can. Pet. Geol.* **14,** 337–81.

—— & Snelson, S. 1980. Realms of subsidence. *In*: *Facts and Principles of World Oil Occurrence.* Mem. Can. Soc. Pet. Geol. **6,** (in press).

Beck, R. H. & Lehner, P. 1974. Oceans, new frontier in exploration. *Bull. Am. Assoc. Petrol. Geol.* **58,** 376–95.

—— & ——, with collab. of Diebold, P., Bakker, G. & Doust, H. 1975. New geophysical data on key problems on global tectonics. *Proc. Ninth World Petrol. Congr. Tokyo,* **2,** Applied Science, Essex, 3–17.

Beloussov, V. V. 1962. *Basic Problems in Geotectonics.* McGraw Hill, New York, 809 pp.

—— 1975. *Foundations of Geotectonics* (in Russian). Nyedra, Moscow, 260 pp.

—— 1977. Gravitational instability and the development of the structure of continents. *In*: Saxena, S. K. & Battacharji, S. (eds). *Energetics of Geological Processes.* Springer, New York, 3–18.

Binns, R. E. 1978. Caledonian Nappe correlation and orogenic history in Scandinavia north of lat. 67°N. *Bull. geol. Soc. Am.* **89,** 1475–90.

Bird, P. 1978. Initiation of intracontinental subduction in the Himalaya. *J. geophys. Res.* **83,** 4975–87.

——, Toksöz, M. N. & Sleep, N. H. 1975. Thermal and mechanical models of continent–continent convergence zones. *J. geophys. Res.* **80,** 4405–16.

Brewer, J., Smithson, S. B., Kaufman, S., & Oliver, J. in press. The Laramide orogeny: evidence from COCORP deep crustal seismic profiles in the Wind River Mountains, Wyoming. *J. geophys. Res.*

Buchanan, R. S. & Johnson, F. K. 1968. Bonanza gas field—a model for Arkoma Basin growth faulting. *In*: Cline, L. M. (ed). *A Guidebook to the Geology of the Western Arkoma Basin and Ouachita Mountains, Oklahoma.* Okla. City Geol. Soc. 75–85.

Buffler, R. T., Shaub, F. T., Worzel, J. L. & Watkins, J. S. 1979. Gravity slide origin for Mexican ridges foldbelt, southwestern Gulf of Mexcio. *Bull. Am. Assoc. Petrol. Geol.,* **63,** 426.

Carey, S. W. 1975. The expanding earth—an essay review. *Earth Sci. Rev.* **11,** 105–43.

—— 1977. The Expanding Earth: *Development in Geotectonics Series,* **10.** Elsevier, Amsterdam, 488 pp.

Cook, F. A., Brown, L. D., Hatcher, R. D., Kaufman, S. & Oliver, J. E. 1979. Preliminary interpretation of COCORP reflection profiles across the Brevard zone in northwest Georgia. *Abstr. Spring Meeting Am. Geophys. Un.*

Curtis, D. M. 1970. Miocene deltaic sedimentation Lousiana Gulf Coast. *In*: Morgan, James P. (ed). *Deltaic Sedimentation Modern and Ancient.* Spec. Publ. Soc. Econ. Paleo. Miner. **15,** 293–308.

—— 1980. Source of Oils in Gulf Goast Tertiary. *J. Sediment. Petrol.* (in press).

—— & Picou, Jr., E. B. 1978. Gulf Coast Cenozoic: A model for the application of stratigraphic concepts to exploration on passive margins. *Trans. Gulf Coast Assoc. geol. Soc.* **XXVIII,** 103–20.

Dahlstrom, C. D. A. 1969. Balanced cross sections. *Can. J. Earth Sci.* **6,** 743–57.

—— 1970. Structural geology in the eastern margin of the Canadian Rocky Mountains. *Bull. Can. Pet. Geol* **18,** 332–406.

Davis, G. A. in press. Problems of intraplate extensional tectonics, Western United States. *In*: *Nat. Res. Council Continental Dynamics publication.*

Davis, G. H. & Coney, P. J. 1979. Geological development of the Cordilleran metamorphic core complexes. *Geology,* **7,** 120–4.

De Charpal, O., Guennoc, P., Montadert, L. & Roberts, D. G. 1978. Rifting, crustal attenuation and subsidence in the Bay of Biscay. *Nature,* **275,** 706–11.

——, Tremolières, P., Jean, F. & Masse, P. 1974. Un example de tectonique de plate-forme: les Causses Majeurs, sud du Massif Central, France. *Rev. Inst. Fr. Pet. Paris* **29,** 641–59.

de Jong, K. A. & Scholten, R. 1973. *Gravity and Tectonics.* Wiley-Interscience, New York, 502 pp.

Delteil, J. R., Rivier, F., Montadert, L., Apostolescu, V., Didier, J., Goslin, M. & Pat-

RIAT, P. H. 1976. Structure and sedimentation of the continental margin of the Gulf of Benin. *In*: DE ALMEIDA, F. F. M., (ed). *Continental Margins of Atlantic Type*. Ann. Acad. Brasil **48,** 51–66.

DE SITTER, L. U. 1954. Gravitation gliding tectonics, an essay in comparative structural geology. *Am. J. Sci.* **252,** 321–44.

EDWARDS, M. B. 1976. Growth faults in Upper Triassic deltaic sediments, Svalbard, *Bull. Am. Assoc. Petrol. Geol.* **60,** 314–55.

EFFIMOFF, I. & PINEZICH, A. A. 1980. Tertiary structural development of selected valleys based on seismic data—Basin and Range Province, Northeastern Nevada. *Paper* for *Roy. Soc. Meeting, London.* (in preparation).

ELLIOTT, D. 1976*a*. The motion of thrust sheets. *J. geophys. Res.* **81,** 949–63.

—— 1976*b*. The energy balance and deformation mechanisms of thrust sheets, *Philos. Trans. R. Soc. London*, **283,** 289–312.

GEE, D. G. 1975. A tectonic model for the central part of the Scandinavian Caledonides. *Am. J. Sci.* **275A,** 468–515.

GIESE, P., MORELLI, C. & STEINMETZ, L. 1973. Main features of crustal structure in central and southern Europe based on data of explosion seismology. *In*: MÜLLER, S. (ed), *Tectonophysics*, **20,** 367–79.

—— & RÜTTER, K. J. 1978. Crustal and structural features of the margins of the Adria Microplate. *In*: CLOSS, H. *et al.* (eds). *Alps, Appennines, Hellenides*, Schweizerbart'sche Verlagsbuchhandlung, Stuttgart, 565–87.

GORDY, P. F., FREY, F. R. & OLLERENSHAW, N. C. 1975. Structural geology of the foothills between Savanna Creek and Panther River SW Alberta, Canada. Guidebook Can. Soc. Petrol. Geol., *Can Soc. Explor. Geophys., Explor. Update* 1975, Calgary.

GWINN, V. E. 1970. Kinematic patterns and estimates of lateral shortening, Valley and Ridge and Great Valley provinces. *In*: FISHER, G. W. *et al.* (eds). *Studies of Appalachian Geology, Central and Southern.* Wiley, New York, 460 pp.

HAMILTON, W. 1977. Subduction in the Indonesian Region. *In*: TALWANI, W. & PITTMAN, W. C. (eds). *Island Arcs. Deep Sea Trenches and Back-arc Basins.* Maurice Ewing Series **1,** Am. geol. Un., 15–31.

HARDING, T. P. 1973. Newport-Inglewood trend, California—an example of wrenching style of deformation. *Bull. Am. Assoc. Petrol. Geol.* **60,** 366–78.

—— 1974. Petroleum traps associated with wrench faults. *Bull. Am. Assoc. Petrol. Geol.* **58,** 1290–304.

—— 1976. Tectonic significance and hydrocarbon trapping consequences of sequential folding synchronous with San Andreas faulting, San Joaquin Valley, California. *Bull. Am. Assoc. Petrol. Geol.* **60,** 366–78.

—— & LOWELL, J. D. 1979. Structural styles, their plate tectonic habitats and hydrocarbon traps in petroleum provinces. *Bull. Am. Assoc. Petrol. Geol.* **63,** 1016–58.

HATCHER, R. D. 1972. Developmental model for the southern Appalachians. *Bull. geol. Soc. Am.* **83,** 2735–60.

HOPKINS, H. R. 1968. Structural interpretations of the Quachita Mountains. *In*: CLINE, L. M. (ed). *A Guidebook to the Geology of Western Arkoma Basin and Ouachita Mountains, Oklahoma.* Okla. City Geol. Soc. 104–8.

HOSSACK, J. R. 1978. The correction of stratigraphic sections for tectonic finite strain in the Bygdin area, Norway. *J. Geol. Soc. London*, **135,** 229–41.

HSÜ, K. J. 1979. Thin skinned plate tectonics during Neo-Alpine Orogenesis *Am. J. Sci.* **279,** 353–66.

HUMPHRIS, C. C. 1978. Salt movement on continental slope, northern Gulf of Mexico. *Am. Assoc. Petrol. Geol. Studies in Geology*, **7,** 69–85.

KOBER, L. 1928. *Der Bau der Erde*, Bornträger, Berlin. 500 pp.

LAUBSCHER, H. P. 1978. Foreland folding. *Tectonophysics*, **47,** 325–37.

LEHNER, P. & de RUITER, P. A. C. 1977. Structural history of the Atlantic margin of Africa. *Bull. Am. Assoc. Petrol. Geol.* **61,** 961–81.

LETOUZEY, J. & TREMOLIÈRES, P. 1980. Paleostress fields around the Mediterranean since the Mesozoic derived from microtectonics: Comparisons with plate tectonic data. *In*: SCHEIDEGGER, A. E. (ed). *Tectonic Stress in the Alpine-Mediterranean region.* Springer. Verlag.

LOWELL, J. D. 1972. Spitsbergen Tertiary orogenic belt and the Spitsbergen fracture zone. *Bull. geol. Soc. Am.* **83,** 3091–102.

MCDONALD, R. E. 1976. Tertiary tectonics and sedimentary rocks along the transition basin and range province to plateau and thrust belt province, Utah. *Rocky Mtn. Assoc. Geol. Guidebook*, 281–371.

MOLNAR, P. & GRAY, D. 1979. Subduction of continental lithosphere: some constraints of certainties. *Geology*, **7,** 58–62.

—— & TAPPONIER, P. 1978. Active tectonics of Tibet. *J. geophys. Res.* **83,** 5361–75.

MUELLER, S. 1977. A new model of the continental crust. *Monogr. Am. Geophys. Un.* **20,** 289–317.

——, EGLOFF, R. & ANSORGE, J. 1976. Struktur des tieferen undergrundes Entlang der Schwizer geotraverse. *Schweiz. mineral. petrogr. Mitt.* **56,** 685–92.

NORTH, F. K. 1964. Gravitational tectonics. *Bull. Can. Pet. Geol.* **12,** 185–225.

OXBURGH, E. Z. 1972. Flake tectonics and continental collision. *Nature*, **239,** 202–4.

PIERCE, W. G. 1957. Heart Mountain and South Fork detachment thrusts of Wyoming. *Bull. Am. Assoc. Petrol. Geol.* **41,** 519–626.

—— 1963. Reef Creek detachment fault, northwestern Wyoming. *Bull. geol. Soc. Am.* **74,** 1225–36.

—— 1973. Principal features of the Heart Mountain fault and the mechanism problem. *In*: DE JONG, K. A. & SCHOLTEN, R. (eds). *Gravity and Tec-*

tonics. Wiley-Interscience, New York, 457–71.

Prey, S. 1974. External zones. *In*: Mahel, M. (ed). *Tectonics of the Carpathian Balkan Regions—Explanation to the Tectonic Map of the Carpathian–Balkan Regions and their Foreland.* Geol. Inst. Dionyz Stur. Bratislava, 75–84.

Price, R. A. 1971. Gravitational sliding and the foreland thrust and fold belt of the North American Cordillera: Discussion. *Bull. geol. Soc. Am.* **77,** 1133–38.

—— 1973. Large-scale gravitational flow of supracrustal rocks, southern Canadian Rockies. *In*: de Jong, K. A. & Scholten, R. (eds). *Gravity and Tectonics.* Wiley-Interscience, New York, 491–502.

—— & Mountjoy, E. W. 1970. Geologic structure of the Canadian Rocky Mountains between Bow and Athabasca Rivers: a progress report. *Spec. Pap. geol. Assoc. Can.* **6,** 7–25.

Reeves, F. 1946. Origin and mechanics of the thrust faults adjacent to the Bear-paw Mountains, Montana. *Bull. geol. Soc. Am.* **57,** 1033–48.

Rider, M. H. 1978. Growth faults in carboniferous of western Ireland, *Bull. Am. Assoc. Petrol. Geol.* **62,** 2191–213.

Roeder, D., Gilbert, Jr., D. E. & Witherspoon, W. D. 1978. Evolution and macroscopic structure of Valley and Ridge thrust belt. Tennessee and Virginia. *Univ. Tenn. Geol. Sci. Studies in Geology,* **2,** Knoxville, 25 pp.

Royse, Jr., F., Warner, M. A. & Reese, D. C. 1975. Thrust belt structural geometry and related stratigraphic problems, Wyoming–Idaho–northern Utah. *Rocky Mtn. Ass. Geol. Symp.* Deep Drilling Frontiers in Central Rocky Mtns., 41–54.

Scholl, D. W., Marlow, M. S. & Cooper, A. K. 1977. Sediment subduction and offscraping at Pacific Margins. *In*: Talwani, M. & Pittman, W. C. (eds). *Island Arcs. Deep Sea Trenches and Back-Arc Basins.* Maurice Ewing Series **1,** Am. geophys. Un., 199–210.

Shouldice, G. H. 1971. Geology of the western Canadian continental shelf. *Bull. Can Pet. Geol.* **19,** 405–36.

Smithson, S. B., Brewer, J. A., Kaufman, S., Oliver, J. & Hurich, C. 1978. Nature of the Wind River thrust, Wyoming, from COCORP deep reflections and gravity data. *Geology,* **6,** 648–52.

——, ——, ——, —— & —— 1979. Structure of the Laramide Wind River uplift, Wyoming, from COCORP deep reflection data and from gravity data. *J. geophys. Res.* **84,** 5955–72.

Spengler, E. 1953-59. Versuch einer Rekostruktion des Ablagerungsraumes der Decken der nördlichen Kalkalpen I-III. *Jahrb. geol. Bundesanst.* 1953, 1956, 1959.

Staub, R. 1928, *Der Bewegungsmechanismus der Erde,* Bornträger, Berlin, 270 pp.

Suppe, J. 1979. Cross Section of southern part of northern coast ranges and Sacramento Valley, California. *Geol. Soc. Am. Map & Chart Series* **MC-28B.**

Sylvester, A. G. & Smith, R. R. 1976. Tectonic transpression and basement-controlled deformation in the San Andreas fault zone, Salton trough, California. *Bull. Am. Assoc. Petrol Geol.* **60,** 2081–102.

Tapponier, P. & Molnar, P. 1976. Slip-line theory and large-scale plate tectonics. *Nature,* **264,** 319–24.

Trümpy, R. 1969. Die Helvetischen Decken der Ostschweiz: Versuch einer palinspastischen Korrelation und Ansätze zu einer kinematischen Analyse. *Eclog. geol. Helv.* **62,** 105–38.

Von Huene, R., Arthur, M. & Carson, B. *this volume.* Ambiguity in interpretation of seismic data from modern convergent margins: an example from the IPOD Japan trench transect.

Wagner, G. H. 1964. Druckspannungsindizien in den Sedimenttafeln des Rheinischen Schildes. *Geol. Rdsch.* **56,** 906–13.

Watkins, J. S., Ladd, J. W., Shaub, F. J., Buffler, R. T. & Worzel, J. L. 1976. Southern Gulf of Mexico, east-west section from Tamaulipas shelf to Campeche Scarp. *Am. Assoc. Petrol. Geol. Seismic sec.* **1.**

——, ——, Buffler, R. T., Shaub, F. G., Houston, M. H. & Worzel, J. L. 1978. Occurrence and evolution of salt in deep Gulf of Mexico. *Am. Assoc. Petrol. Geol. Studies in Geology,* **7,** 43–65.

Weber, K. J. & Daukoru, E. 1976. Petroleum geology of the Niger delta. *Proc. Ninth World Petrol. Congr. Tokyo,* **2,** Applied Science, Essex, 209–221.

Wells, A. T., Forman, D. J., Ranford, L. C. & Cook, P. J. 1970. Geology of the Amadeus Basin, central Australia. *Bull. Bur. Miner. Resour. Geol. Geophys. Melbourne.* **100,** 222 pp.

Wilcox, R. E., Harding, T. P. & Seely, D. R. 1973. Basic wrench tectonics. *Bull. Am. Assoc. Petrol. Geol.* **57,** 74–90.

Wunderlich, H. G. 1974. Die Bedeutung der Süddeutschen Gross-scholle in der Geodynamik Westeuropas. *Geol. Rdsch.* **63,** 755–72.

Ziegler, P. A. 1978. Northwestern Europe: tectonics and basin development. *Geol. Mijnbouw.* **57,** 589–626.

A. W. Bally, Shell Oil Company, P.O. Box 481, Houston, Texas 77001, U.S.A.