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NORMAL FAULTING, BLOCK TILTING,
AND DÉCOLLEMENT IN A STRETCHED CRUST

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Abstract. Extensional tectonics cover a wide range of crustal structures from narrow continental rifts to large continental areas. If we except continental and oceanic rifts, stretched portions of the crust exhibit complex faulting patterns at upper levels with high-angle and low-angle normal faults. Very little is known concerning the geometry of structures and physical processes at depth. It is therefore difficult to extrapolate surface observations to the crustal scale. Five general models are proposed that combine continuous and discontinuous deformation and geometrical and geological tests are proposed. Two natural examples, the West Armoricain Atlantic margin and Eldorado Mountains (Basin and Range Province), are presented. They are characterized by an association of high-angle normal faults, tilted blocks, and low-angle normal faults. Field and geometrical arguments are given to demonstrate that low-angle faults are controlled by décollement surfaces along preexisting interfaces. Geometrical implications of the tilted block pattern are presented and the gravity gliding hypothesis is proposed as a possible explanation of block tilting when associated with low angle normal faults. In such

cases, the amount of stretching cannot be related to bulk thinning of the crust.

INTRODUCTION

Most extensional tectonic domains exhibit normal faulting and block tilting at upper levels in the crust. This is commonly observed on passive continental margins or over wide regions such as the Basin and Range Province. Because very little is known about the geometry of structures and physical processes at depth, it is difficult to extrapolate surface observations to the crustal scale. The purpose of this paper is to examine, from general considerations and natural examples, whether one can use surface structures to estimate the amount of stretching.

We first propose five general models of stretched crust and discuss their specificities. Secondly, two natural examples are considered, the West Armoricain Atlantic Margin and the Basin and Range Province of Western America. We give geological and structural arguments that lead us to interpret low-angle normal faults in these examples as décollement surfaces controlled by flat-lying preexisting interfaces. Lastly, we discuss field and geometric tests of the models. The geometrical implications of tilted block patterns are presented with some details. The gravity gliding hypothesis is proposed as a possible explanation of block tilting when associated with low-angle normal faults.

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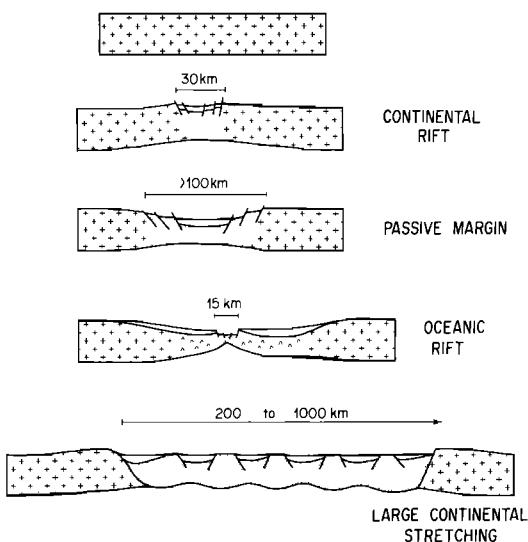


Fig. 1. The main structural occurrences of stretching in the crust. The instantaneous deformed zone is dotted.

STRETCHING OF THE CRUST

The main types of stretched portions of the crust are (1) continental rifts, which are generally considered as early stages of stretching in the continental crust, (2) continental margins, as a stage following the previous and preceding, (3) oceanic rifts, and (4) large continental areas such as the Basin and Range Province and the Tibetan Plateau.

The striking features that result from this classification (Figure 1) is the scale variation from one to another. The width of active area during stretching in the four types of ranges from 15 to 1000 km. Implied volumes of crustal material are therefore dramatically different, and physical implications should not be equivalent. For example, the balance between boundary forces and body forces must vary from one type to the other. From the structural point of view, it is interesting to note that in those types for which the instantaneous stretching domain is narrow (oceanic and continental rifts), low-angle normal faults have never been observed.

STRUCTURAL MODELS

Five structural models of stretched crust may be proposed if we take into account the following parameters:

1. The vertical zonation into domains in which deformation is mainly achieved either

by discontinuous (ductile) or discontinuous (brittle) processes, and the nature of the transition between the two types of domains.

2. The nature of discontinuities in the "discontinuous domain."
3. The presence or lack of preexisting or inherited low-dipping mechanical interfaces.
4. The existence of a gradient of stretching across the zone.

The five models are (Figure 2) as follows:
Model I. The crust is thinned only by a purely ductile process (Figure 2a). This model is probably fictitious.

Model II. A progressive vertical transition from purely brittle deformation at the top to purely ductile deformation at the base is supposed (Figure 2b). In this case, there is a direct relation between the amount of stretching achieved by normal

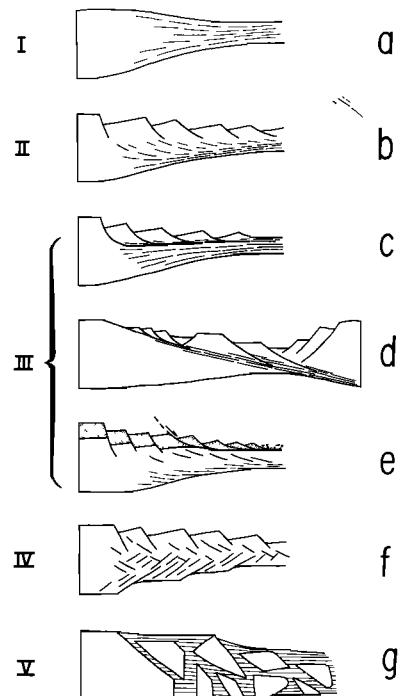


Fig. 2. The five models of stretched crust obtained by a combination of continuous and discontinuous processes: (a) completely continuous; (b) progressive transition between continuous and discontinuous; cases where a preexisting mechanical interface localizes: (c) the boundary between the two domains, (d) the ductile deformation, and (e) the sliding of an upper plate; (f) completely discontinuous; (g) completely discontinuous with addition of material.

faults and block tilting at the surface and the bulk thinning of the crust. Such a model is supported by the fact that many field studies tend to demonstrate that faults pass into ductile shear zones at depth [see Sibson, 1977].

Model III. In this model, we consider the presence of mechanical interfaces. Three types of interfaces may be implied, leading to three subclasses of models (Figures 2c, 2d and 2e).

If a strong temperature gradient can occur in the crust, one can suppose that a thin transition zone may separate an upper brittle layer from a lower ductile layer (Figure 2c). This hypothesis can be based on the fact that rheological properties of crustal materials are strongly temperature dependent. Such a model is implicitly presented by Coletta and Angelier [1982] and Angelier and Coletta [1983] and has been studied in detail from the structural and geophysical points of view by Profett [1977], Montadert et al. [1979], Le Pichon and Sibuet [1981] and Le Pichon et al. [1982]. Because no slip occurs at the interface, block tilting at the surface reflects the amount of stretching and is directly related to the thinning of the crust.

If stretching is able to induce a low angle normal fault at the scale of the crust, deformation may be summarized as the combination of a low dipping ductile shear zone and block faulting above it (Figure 2d). In this model, proposed by Wernicke [1981], high-angle normal faults and associated tilted blocks may be used to estimate a minimum value for the amount of stretching.

If the crust contains preexisting mechanical interfaces such as lithological discontinuities, basement-cover boundary, or ancient thrust zones, a vertical decoupling can occur on them, especially if body forces are able to become active (Figure 2e). Low-angle normal faults are related to gravity sliding structures [Hose and Danes, 1973; Davis et al., 1980] and eventually controlled by preexisting interfaces. Block faulting that develops to accommodate the sliding is not representative of the amount of stretching undergone by the crust, as the blocks may be destroyed and resedimented at the front of the glided slab [Embley, 1976; Jorgensen et al., 1982]. In this model, an estimate of crustal stretching made from tilted blocks at the surface should be generally overestimated.

Model IV. The deformation may be purely

discontinuous (Figure 2f). One can expect an increase of fracturation density at depth.

Model V. The stretching may be accommodated by intrusion of magmas during the process. Structures in the blocks give an amount of stretching lower than the true stretching.

Main features of these models are summarized on Table 1. Other models may be obtained by combination of these elementary ones.

ANALYSIS OF NATURAL EXAMPLES

If we except the case of oceanic and continental rifts, most stretched areas show at the surface an association, in time and space, of high-angle normal faults (HANF's) bounding tilted blocks and low-angle normal faults (LANF's). Some authors have discussed the possibility of using the geometry of faults and tilted blocks to estimate the amount of stretching [Le Pichon and Sibuet, 1981; Wernicke and Burchfield, 1982]. Such estimates vary as to whether one considers LANF's to be the ultimate evolution of HANF's [Le Pichon and Sibuet, 1981; Coletta and Angelier, 1982] or, more generally, if LANF's are interpreted by another model LANF's [e.g., Wernicke, 1981]. Geological evidence for estimating the amount of stretching is discussed by using two natural examples:

The West-Armorian passive margin. The West-Armorian passive margin has been recently proposed as a typical example in which the stretching estimated from block tilting (up to 300 % is in good agreement with the thickness of the crust (12 km) [Le Pichon and Sibuet, 1981]. However, if we look at seismic profiles and deduced geological cross sections [Guenno, 1978; Roberts and Montadert, 1980; Le Pichon and Sibuet, 1981], we note that discordant lower Cretaceous to Tertiary basins are situated between the tilted blocks. Another striking feature is the fact that the LANF's that separate the blocks seem to join a seismic velocity discontinuity between a 4.9 km s^{-1} layer and a 6.3 km s^{-1} layer at depth (Figure 3). The following facts must be added: (1) in other available cross sections of the margin, tilted blocks are not always present, (2) the existence, in some places, of undisturbed reflectors below the seismic discontinuity (Société Nationale Elf Aquitaine Production, personnel communication, 1983), (3) laterally, in a more eastern part of the margin,

TABLE 1. Main Features of the Five Models (See Figure 2)

Models	I	II	III	IV	V	
Reference to Figure 2	a	b	c	d	e	f
Existence of seismic reflectors	0		0	0	0	
Rheological contrast			0			
Destruction and resedimentation of material					0	
Existence of low-angle normal faults			0	0	0	0
Ductile behavior at the base of the crust	0	0	0	0	•	
Good agreement between stretching at the surface and thickness of the crust	0	0	0	0		0

Open circles mean yes, solid circles mean yes or no.

where the discontinuity is observable at the same level, salt diapirs are initiated above it (Société Nationale Elf Aquitaine Production, personnel communication, 1983), (4) the Jurassic has been reached by the 48 IPOD drill at the top of Meriadzek tilted block, (5) granitic and paleozoic material has been dredged in the northern part of the tilted blocks area [Auffret et al., 1979].

Taking into account these facts, we assume that (1) the tilted blocks are mainly made up of jurassic and triasic material, (2) the seismic discontinuity is probably the mesozoic cover/hercynian basement boundary as interpreted by Guennoc [1978], (3) this boundary has been the locus of a local decollement of the mesozoic cover affected by block faulting and so translated southward during lower cretaceous time, (4) the northern outcrops of hercynian basement represent a denuded zone as a result of this translation, (5) at least, this observation is certainly local and cannot be generalized because we know that, at many other places, HANF's affect the basement itself and that horst and graben structures have been recognized [Guennoc, 1978]. It is, however, very important to point out that at these places, the stretching estimated

from block faulting is considerably lower (30-40 %; Chenet [1982]) than those estimated by Le Pichon and Sibuet [1981]. Moreover, structural maps of this passive margin [Guennoc, 1978] show that there is no simple evolution of fault dips indicating a continuous increase of stretching from the undeformed continental crust to the beginning of oceanic crust.

The evidence for tilted blocks above a lithological interface is not uncommon. On the Northern Mediterranean margin, the Messinian salt layer separates recent sediments from their Meso to Cenozoic substratum. Here again, normal faults in the upper layer join the Messinian salt layer at depth and older terrains show a horst and graben structure (Biju-Duval et al. [1974]; see also Humphris [1978] for the same phenomena occurring in the Gulf of Mexico). Another example is given by Dingle [1977, 1980] in the margin of South Africa where the lithological boundary between Paleogene and Cretaceous and the base of Neogene sediments controls the sliding of these units over their substratum. Decoupled units may have undergone large extension, and their front has been observed layed over the oceanic crust and partially resedimented [Embley, 1976].

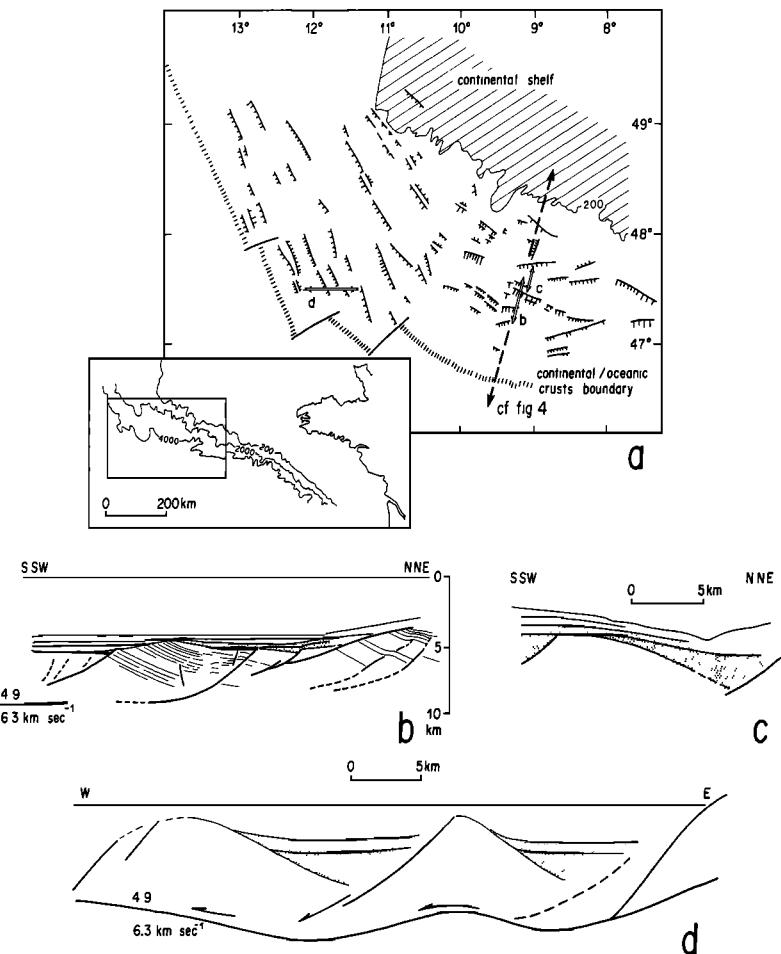


Fig. 3. The West-Armorian passive margin [after Guennoc, 1978; Montadert et al., 1979]: (a) location and map of normal faults; seismic cross sections located on Figures 3a, 3b, and 3c are migrated, Figure 3d is nonmigrated. Dotted layers represent probably Albian and Aptian sediments. Note the curvature of the normal faults, delimiting tilted blocks, and joining the boundary between 4.9 and 6.3 km/s layers.

These facts show that on passive margins (1) a preexisting mechanical interface can induce and control the development of a stretching fault pattern above it, and (2) the vertical offset along individual HANF's can cumulate at the interface and create a décollement (LANF) (Figure 4).

The Eldorado Mountain-Basin and Range Province. Four types of faults and discontinuities have been described in the Basin and Range Province (Figure 5): (1) HANF's delimitating recent sedimentary basins (Miocene to Present), giving the basin and range morphology to the region; the vertical displacement along these faults may be up to 4 km [Effimoff and Pinezich, 1981; Anderson et al., 1982]; (2) LANF's of crustal dimension, either outcropping, or recog-

nized on seismic profiles over large distances and down to 20 km at depth [Anderson et al., 1982]; (3) strike slip faults whose role is probably to accommodate differences in stretching from different part of the Basin and Range Province; (4) listric normal faults whose direction is parallel to those of HANF's. Their dip varies from 50 to 0° giving to the blocks a lenticular aspect. The tilting of the blocks can be as high as 90°.

The last type of faults has been recognized in the Eldorado Mountains by Anderson [1971], who has introduced the concept of thin-skin distension to describe the phenomena. Since Anderson, numerous occurrences of this fault pattern have been described [Profett, 1977; Davis et al.,

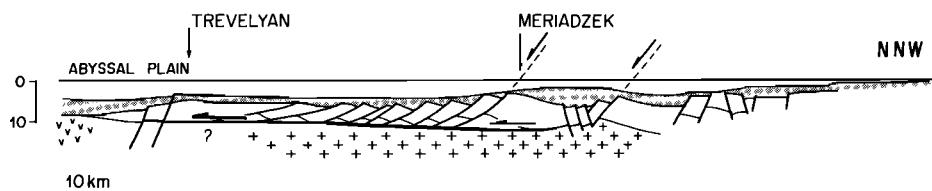


Fig. 4. Interpretative cross section of the West Armorican passive margin (see location on Figure 3). The Mesozoic pre-Albian formation, where tilted blocks occur, has suffered a displacement along the Hercynian basement/sedimentary cover boundary.

1980; Gross and Hillemeyer, 1982; Mathis, 1982; Spencer, 1982], and their importance is widely recognized. They affect Miocene volcanic series and appear to have been active during a short time interval (about 10 M.Y.). Excellent field arguments show that they were active before basin formation but compatible with the same stress field [Zoback et al., 1981]. Most authors consider that these faults reflect directly the stretching undergone by volcanics and sediments during the Basin and Range dis-tension.

Recent field work by one of us (P.C.) shows that two types of preexisting mechanical interfaces are involved in the stretching:

1. In the volcanic series substratum. They are mainly ancient thrust zones of Cordilleran age (see geological map of Nevada [Anderson, 1977, 1978]) easily observable in Precambrian and Paleozoic rocks. In the former, mylonitic zones developed in older amphibolitic facies rocks have been interpreted as the ductile equivalent at depth of the brittle deformation observed in the upper layers [Davis and Coney, 1979; Wernicke, 1981]. Recent work in the Whipple Mountains [Davis et al., 1980], in the Riverside Mountains [Lyle, 1982] and in the Eldorado region (P. Choukroune and E. Smith, manuscript in preparation, 1983) has de-

monstrated that these large mylonitic zones were related to an older event, without kinematic relationship with the late stretching. Nevertheless, they introduce important discontinuities in the crust. As far as we can judge, the major LANF's described by Wernicke [1981] and Anderson et al. [1982] resemble this type of preexisting discontinuities.

2. The basal boundary of the volcanic series and more generally all important lithologic boundaries in these series constitute the second type of preexisting discontinuity. In the Eldorado Mountains, we can observe (1) that the volcanics at the map scale [see Anderson, 1978] are separated from the Precambrian basement by a décollement surface and (2) at the outcrop scale that the lithological discontinuities in the volcanics show traces of movement. Moreover, curved faults join very frequently these discontinuities at all scales (Figure 6).

To these data, it is important to add the following facts:

1. Everywhere in the Eldorado Mountains, the deformation that can be related to the stretching is purely brittle. From the base of the Cenozoic series (4–5 km thick) to its top, there is no variation of deformation mechanisms.

2. The displacement of the series involved

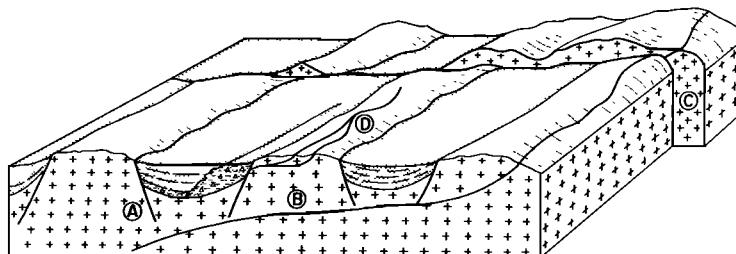


Fig. 5. Schematic block diagram of fault types in the Basin and Range Province. (a) high-angle normal faults delimitating the basins; (b) major low-angle normal faults seen on seismic profiles; (c) strike-slip faults; (d) low-angle normal faults delimitating tilted blocks.

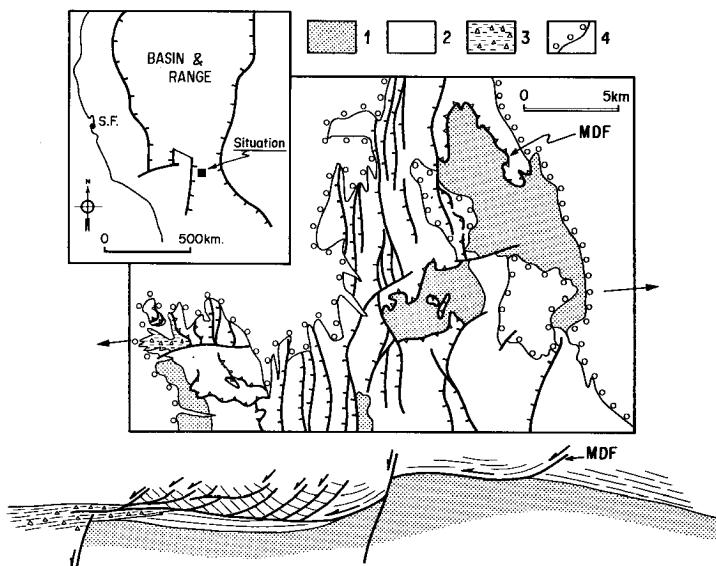


Fig. 6. Map and section across the Eldorado Mountains (Basin and Range Province, Nevada). M.D.F.: Main Décollement Fault. 1, Precambrian rocks. 2, Tertiary volcanics. 3, detrital continental material. 4, postdécollement deposits. Map simplified after Anderson [1978]. The late normal faulting is not represented on the cross section.

in the décollement is always directed toward the W-NW.

3. Fracture density increases from east to west; that is to say, the dimension of tilted blocks decreases toward the west.

4. At their front, the glided series are extremely jagged and lie over conglomeratic series or are enclosed by them ("structural chaos" of Anderson [1978]) (Figure 6) (see also Jorgensen et al. [1982]).

DISCUSSION

Tilted Blocks Patterns

In many areas subjected to stretching, the striking structural feature at the surface is the tilted blocks pattern. We illustrated some geological and structural aspects of this pattern for the Basin and Range Province and the West Armorican passive margin. Some authors [Le Pichon and Sibuet, 1981; Wernicke and Burchfiel, 1982] have proposed that geometrical relations between faults and bedding may be used to obtain an estimation of the amount of stretching, if one suppose that the blocks remain rigid during the deformation.

Let us consider a system of tilted rigid blocks separated by parallel plane faults, such that the line joining the top apices of the blocks is horizontal (Figure 7). Before block tilting, the fault dip is p_0 and the fault spacing is equal to 1. As stret-

ching evolves, the resulting fault dip p decrease ($p < p_0$) and the tilt of the block increases as

$$\alpha = p_0 - p \quad (1)$$

The corresponding horizontal stretch e can be expressed as a function of initial and finite fault dip:

$$\frac{(1 + e)}{\sin p_0} = \frac{1}{\sin p} \quad (2)$$

This relation is equivalent of equation (8) in the work of Le Pichon and Sibuet [1981] and formulae in Figure 2 by Wernicke and Burchfiel [1982]. Figure 7 shows the variation of p and α as functions of $(1 + e)$ for different initial values of p_0 and α_0 . We note that (1) rapid variations (significant) of fault dip occur during the early 50% stretching (i.e., $(1 + e) = 1.5$); (2) faults with initial dip ranging between 90° and 50° (i.e., a 40° fan) give at a 100% stretching a 7° fan; (3) horizontal faults, whatever their initial dip if not null, are only expected for an infinite amount of stretching; (4) the amount of tilting becomes rapidly equal to the fault dip; for example, if $p_0 = 60^\circ$, α is equal to p ($\alpha = p = 30^\circ$) for $(1 + e) = 1.8$ (i.e., a 80% stretching); (5) high amounts of tilting can be obtained only if initial values of p_0 are high.

If the blocks do not remain rigid during

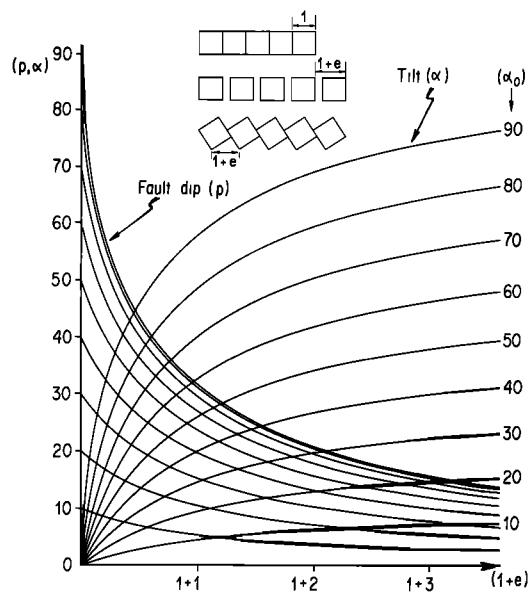


Fig. 7. Fault dip p and block tilt α variations as a function of finite stretch $(1 + e)$ where e is $\Delta l/l$. Initial values of p (p_0) and α (α_0) are indicated.

deformation, the model is no longer valid. A qualitative model of structural evolution for deformable blocks has been proposed recently by Coletta and Angelier [1982] and Angelier and Coletta [1983] (Figure 8). These authors (1) observe that block tilting is accommodated by second- and third-order faults perpendicular to bedding in the blocks and (2) quote that internal deformation remains negligible until stretching reaches values up to 50–100%. Because it is also implied that strain increases downward, faults that are initially flat progressively become curved during block tilting. The model is able to give a unique explanation of observed fault patterns whatever the associated fault dip and bedding tilt. We believe important to note that in this deformable block model, (1) finite strain gradient is necessarily associated with first-order faults, and (2) low-angle normal faults result from progressive evolution of faults having an initially greater dip.

The reasons why tilted block models have been applied to continental margins [Le Pichon and Sibuet, 1981] and Basin and Range [Coletta and Angelier, 1982] are obvious. But according to us, this may be questioned because (1) 90° tilts associated with quasi-horizontal faults (e.g., Basin and Range) imply very high amounts of stretching, and (2) even if the deformable

block model can reconcile this aberration, finite strain gradients are not observed near the first-order faults.

Moreover, in the two examples, it is dubious that low-angle normal faults result from the evolution of faults having initially higher dips.

Significance of low-angle normal faults.

Considering the lower boundary of the faulted layer, two end member boundary conditions are possible (Figure 9):

1. Sliding is allowed at the base of the blocks (i.e., the lower boundary is an incoherent interface).
2. Relative movements of the blocks are accommodated by continuous deformation under the faulted layer (i.e., the lower boundary is a coherent interface between an upper brittle layer and a lower ductile layer).

In case 2, the tilted block pattern can be used to estimate the amount of stretching undergone by the system, but the base of the blocks cannot be considered as a fault.

Field work in the Basin and Range Province has shown that tilted blocks are bounded downward by two types of low-angle normal faults under which the deformation due to stretching is not ductile. The first type of LANF, the smaller (surface A on Figure 10a), is associated witholistostromes in the basins and probably correspond to megalandslides. These faults are controlled by the bedding anisotropy in sedimentary and volcanic rocks, or by the cover–basement boundary. In this case, the tilted blocks overestimate the amount of true crustal stretching. The second type, the larger, can be observed on seismic profiles (surface B on Figure 10a). These LANF can be interpreted as a normal fault sense reactivation of ancient thrust faults of Cordilleran age (they are interpreted by

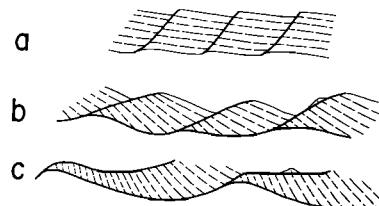


Fig. 8. The progressive evolution (a \rightarrow b \rightarrow c) of deformable tilted blocks (simplified after Coletta and Angelier [1982]). Dashed lines indicate bedding. Second- and third-order faults are not represented.

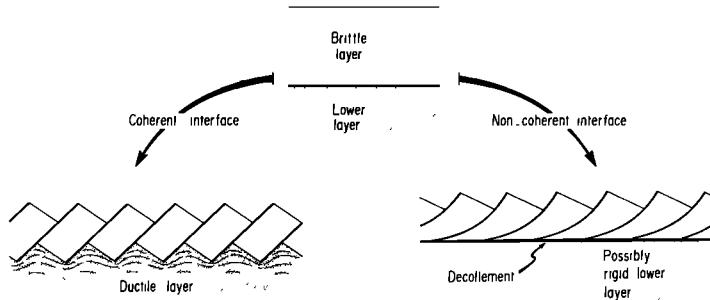


Fig. 9. The two extreme mechanisms of tilting in an upper brittle layer as a function of slip condition at the interface with its substratum.

Wernicke [1981] as major normal faults due to stretching). In this case, the tilted blocks underestimate the amount of true crustal stretching. One can note that this interpretation is in a complete agreement with Davis et al. [1980, 1982].

For passive margins (Figure 10b), it is not possible to verify if ductile deformation has occurred under low-angle normal faults. Arguments have been presented showing that tilted blocks occur over a lithological interface: the cover-basement boundary for the West Armoricain margin and the Messinian salt layer for the Northern Mediterranean margin; other interfaces are reactivated in the case of Gulf of Mexico and the margin of SE Africa.

In the two types of geological situations, the lower boundary of the tilted blocks is an incoherent interface, namely, a low-angle normal fault. A simple model for this situation is the slip line field of an half Prandtl cell [Nadaï 1950, p. 533; Kanizai, 1962] for the extensive flow case [Nye, 1951, 1952]. The slip lines, which correspond to maximum shear stress trajectories, are arranged in two mutually orthogonal sets of cycloids intersecting the free boundary at 45°. In truly plastic materials, faults initiate parallel to the set of slip lines corresponding to extensive flow. Because real rocks are not truly plastic, the model may be used only in a qualitative way. Some refinements are necessary to fit more accurately the fault pattern observed in nature. Notably, the model should explain faults intersecting the upper free boundary at 60°. We believe that such a model is applicable to many situations at different scales in the Basin and Range (previously) proposed by Hose and Danes [1973] and to high stretching tilted blocks in passive margin.

Brittle-Ductile Transition at Depth

It is well known that rock ductility is temperature dependant. Because temperature increases with depth, it is reasonable to expect an increase in rock ductility toward the deeper parts of the crust. But very little is known concerning the level and nature of the brittle-ductile transition. The first problem is to define the ductility. As structural geologists, we use this term to describe the property of a rock to undergo large strains without rupturing. In most crustal rocks, a penetrative cleavage begins to develop when the finite shortening attains 20–25%. Cleavage is one of the striking features of ductile deformation, and mapping of the appearance of cleavage (boundary known as "cleavage front," cf. Fourmarier in the work of Baer [1956] and

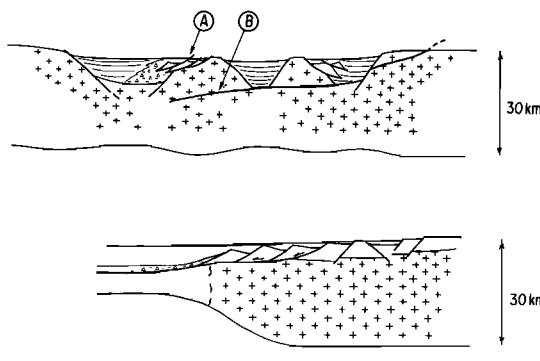


Fig. 10. Synthetic comparative cross sections of Basin and Range (up) and West Armoricain passive margin (down) showing the place and the role of gravity slides. In the Basin and Range Province, they are controlled (a) by lithological interfaces at upper level and (b) by old Cordilleran thrust faults.

Mattauer [1973] helps to define the upper ductile boundary in the field.

Numerous studies in mountain belts have demonstrated that (1) if an upper and external brittle domain can be defined in which no ductile deformation occurs, discontinuities are always present in the ductile domain even at the deepest parts of the crust, (2) the transition between the two domains is always progressive in space and depends on rock composition, and temperature. Large ductile strains occur in pelites from 250°C onward when solution transfer is active and in quartzites from 350°C as a result of the onset of dynamic recrystallization [Gapais and Le Corre, 1981]. In typical basement rocks such as granites, the onset of ductile deformation is characterized by the development of complex networks of ductile shear zones enclosing more or less undeformed blocks [Ramsay and Allison, 1979; Choukroune and Gapais, 1983]. The metamorphic environment of this type of deformation is commonly referred to as the base of greenschist facies or the top of amphibolite facies, i.e., between 400 and 500°C. These field observations are consistent with experimental data [Goetze, 1971; Carter et al., 1981; Alm, 1979] and electron microscopy studies [Vidal et al., 1980]. Taking a temperature gradient of 25°C km⁻¹ as a maximum value preceding the stretching, we can place the upper ductile boundaries using the reference temperatures quoted above between 10–12 km for sediments and at 18 km for a granitic basement. After a homogeneous stretching of 100%, these became 5–6 km for the sediments and 9 km for the basement. Comparing with a natural case such as the West Armoric Margin (Figure 4), we note paradoxically that the Jurassic sediments could have been ductile and the top of the basement brittle. Let us recall that previous authors [Montadert et al., 1979; Le Pichon and Sibuet, 1981] consider the cover as entirely brittle and the basement entirely ductile. As a consequence, it is not surprising (1) that the brittle-ductile transition has not been observed in Basin and Range Province where a 4000-m thickness of the crust is observable and (2) that all the deformation is associated with décollements at preexisting interfaces. In such terranes, the brittle-ductile transition must be seated at greater depths, if it does exist.

Finally, it seems unlikely to us that a heterogeneous system can give a homogeneous deformation. Vertical variations of the amount of stretching should be expected.

CONCLUSIONS

Our main conclusions are as follows:

1. In large domains of instantaneous stretching, preexisting mechanical interfaces (bedding, cover-basement boundary, ancient major faults) play an important role in the development and distribution of new structural discontinuities;
2. The role of body forces must not be neglected as they can be responsible for large gravity slides over preexisting interfaces;
3. Significant discontinuities must be distinguished from "second-order" gravity sliding structures; in other terms, the bulk fault pattern results from a complex sequence of faulting in which some faults are not directly related to the stretching.
4. During stretching deformation, the volume of deformed material may vary as a result of combined destruction and addition of crustal material.
5. Because temperature gradually increases with depth, it is reasonable to expect a proportional increase of rocks ductility. However, ductile deformation due to stretching at depth has never been observed in the continental crust, and we may therefore only speculate on this matter.

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