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Narrow rifts versus wide rifts: inferences for the mechanics of rifting from laboratory experiments

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THE ROYAL SOCIETY

Narrow rifts versus wide rifts: inferences for the mechanics of rifting from laboratory experiments

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Laboratory experiments on analogue models of the lithosphere are useful tools to study tectonic processes and, in particular, to test physical hypotheses. They complement numerical modelling because the inherent limitations of each method are different. The basic principles of the method are recalled with particular application to models simulating the brittle–ductile layering of continental crust and lithosphere using sand and silicone putties to simulate the frictional and viscous behaviour of rocks. A selection of experiments is used to examine the role of rheology on the development of crustal-scale extensional structures: continental rifts, passive margins, wide extended domains, and core complexes. The difference between narrow rifts and wide rifts is attributed to the type of mechanical instability that can develop for a given type of lithospheric strength profile: namely necking versus spreading. Necking occurs preferentially in a stable lithosphere that has a four-layer-type of strength profile with the greatest strength located in the sub-Moho mantle. It gives birth to narrow rifts, from continental rifts to passive margins. Spreading occurs preferentially in a thickened lithosphere whose strength profile, after thermal relaxation, exhibits maximum strength at the base of upper brittle crust. It gives birth to wide rifts, such as the Basin and Range of the western United States or the Aegean. Core complexes are not considered to represent a particular mode of extension but are anomalies in wide rifts.

Keywords: lithosphere extension; analogue models; brittle–ductile coupling; necking; gravity spreading

1. Introduction

Continental lithosphere extension occurs in various plate tectonic environments, but from a structural point of view, a major difference exists between localized rifting within a normal thickness crust (30–40 km) and widespread extension of thickened crust (Brun & Choukroune 1983; Buck 1991). The former, called *narrow rifting*, starts with continental rifts—e.g. Rhine, Baikal or Ethiopian Rifts—and ends up with passive margins and continental break-up. The latter, called *wide rifting*, occurs during (e.g. Tibet, see Armijo *et al*. (1986)) or after the cessation of convergence (e.g. Basin and Range of the western United States, see Coney & Harms (1984)).

The bulk mechanical behaviour of an extending lithosphere is suitably studied using numerical models that incorporate the strong temperature dependence of rock rheology (e.g. Dunbar & Sawyer 1989; Chéry *et al*. 1990; Govers & Wortel 1995). Until now, these models failed to simulate faulting in brittle layers. Laboratory experiments on small-scale models made of brittle and ductile analogue materials

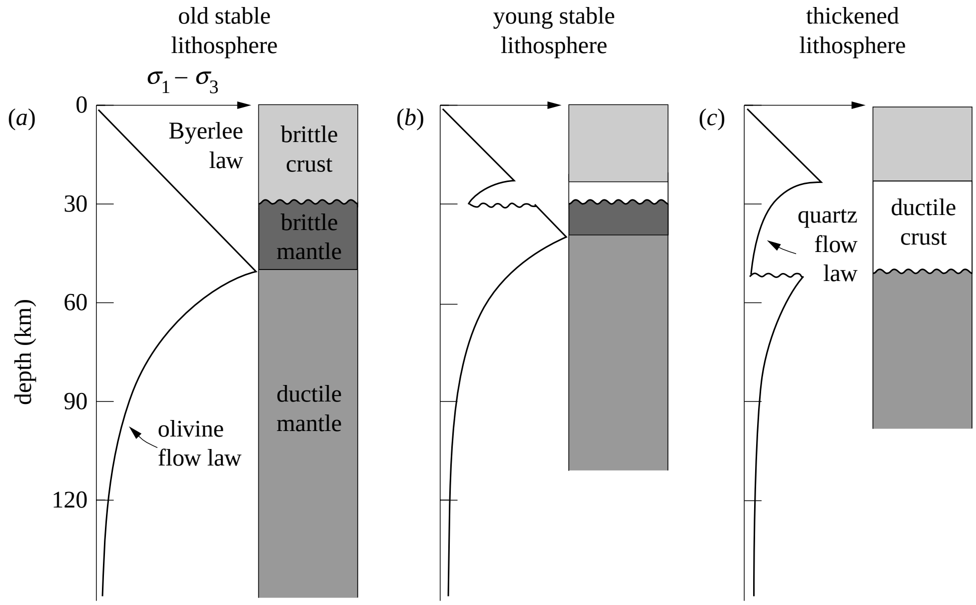

Figure 1. Types of lithosphere strength profile as a function of age and crustal thickness. Lithosphere stabilized (*a*) in pre-Cambrian times; (*b*) in Phanerozoic times; and (*c*) lithosphere still thick and hot.

are unable to suitably incorporate the temperature dependence of rock rheology but were proved successful for the study of both bulk and internal mechanical instabilities of an extending lithosphere (e.g. Faugère & Brun 1984; Allemand *et al.* 1989; Vendeville *et al.* 1987; Brun *et al.* 1994; Brun & Beslier 1996; Benes & Davy 1996).

The present paper reviews more than 15 years of the modelling of extensional tectonics in the experimental tectonics laboratory of Geosciences Rennes, using the sand–silicone multilayer technique. A selection of significant results is used to present the sensitivity of sand–silicone multilayers to variations in brittle and ductile strengths and their effects on the initiation and development of extensional structures. Analogue models are compared with natural examples to examine the mechanical and structural consequences of various types of strength profiles and the differences between narrow rifts and wide rifts, including core complexes.

2. Narrow rifts versus wide rifts

In a stabilized continental lithosphere undergoing extension with a temperature of *ca.* 600 °C at the Moho, at least two strength peaks occur, one below the Moho and the other in the middle crust, separated by a low-strength lower crust (figure 1*b*). Even though one cannot exclude the possibility that a weak layer is no longer present in the lower crust of an old and cold lithosphere stabilized during the Precambrian that has experienced a fast strain rate (figure 1*a*), there is general agreement to use four-layer-type strength models to mechanically model narrow rifting both in numerical simulations and in analogue experiments. In a thickened lithosphere, the strength of the sub-Moho mantle is greatly reduced giving, 20 Ma after the cessation of thickening, a dominant strength peak in the middle crust at the base of the upper brittle crust (figure 1*c*). Therefore, mechanical modelling of wide rifting generally refers to two- or three-layer-type strength profiles.

(*a*) *Narrow rifts*

Lithosphere extension results in a necking-type instability when the geotherm is normal and the crustal thickness is *ca.* 30–40 km (figure 2*a*). In geological terms, continental rifts represent the early stage of necking. They have characteristic widths of 30–40 km at the onset of necking but some exceptions exist (e.g. 60–70 km in the East African Rift). Below continental rifts, the Moho is bent upward but with a wavelength longer than the width of the rift itself. Passive margins represent the later stage of necking. Their characteristic width commonly ranges between 100 and 400 km. Crustal thickness decreases from 30–40 km landward to *ca.* 8–10 km at the continent–ocean boundary. Steeply dipping normal faults dominate landward, producing large slightly tilted blocks or horsts and grabens. Approaching the continent–ocean boundary, blocks become smaller, thinner and more strongly tilted, and faults dip at lower angles.

Although the present paper is dedicated to continental lithosphere rifting, it is interesting to recall, here, that oceanic sea-floor spreading is accommodated within narrow rifts whose characteristic widths range between 1 km for fast-spreading rates (e.g. East Pacific Rise) and 10 km, or more for slow-spreading rates (e.g. Mid-Atlantic Ridge; see Choukroune *et al*. 1984). When compared to continental rifts, oceanic rifts are probably better explained in terms of steady-state necking (Tapponnier & Francheteau 1978).

(*b*) *Wide rifts*

In wide extended domains, such as the Basin and Range of the western United States or the Aegean, extension follows the cessation of a previous period of crustal thickening during continental collision by more than 20 Ma, a long enough time for thermal relaxation to have considerably weakened the thickened crust and the underlying mantle (Sonder *et al*. 1987). The resulting strength profiles have their major peak located at the base of the upper brittle crust (figure 2*b*). The lithosphere is then able to spread under its own weight. The width of wide rifts can be as large as the previously thickened domain amplified by the amount of extension, and can reach 1000 km, as observed in the Basin and Range as well as in the Aegean. In both examples, crustal thickness returned to *ca.* 30 km during extension. Wide rifts are characterized by horsts and grabens, or by tilted blocks arranged in domains of variable vergence, and by the so-called metamorphic core complexes, which correspond to local zones of exhumed ductile lower crust developed at early stages of extension.

Flat-lying detachment faults associated with core complexes were initially interpreted as normal faults that had initiated at a low angle and had cross-cut the whole lithosphere (Wernicke 1985). However, at variance with narrow rifts, seismic reflection profiling in the Basin and Range demonstrated that the Moho is flat below the whole extended domain and shows no sign of strong normal offset in the vicinity of core complexes (Allmendinger *et al*. 1987). More recent analyses (Buck 1988; Wernicke & Axen 1988) argued that detachment faults can originate from forward sequences of steeply dipping normal faults progressively rotated to very low dip.

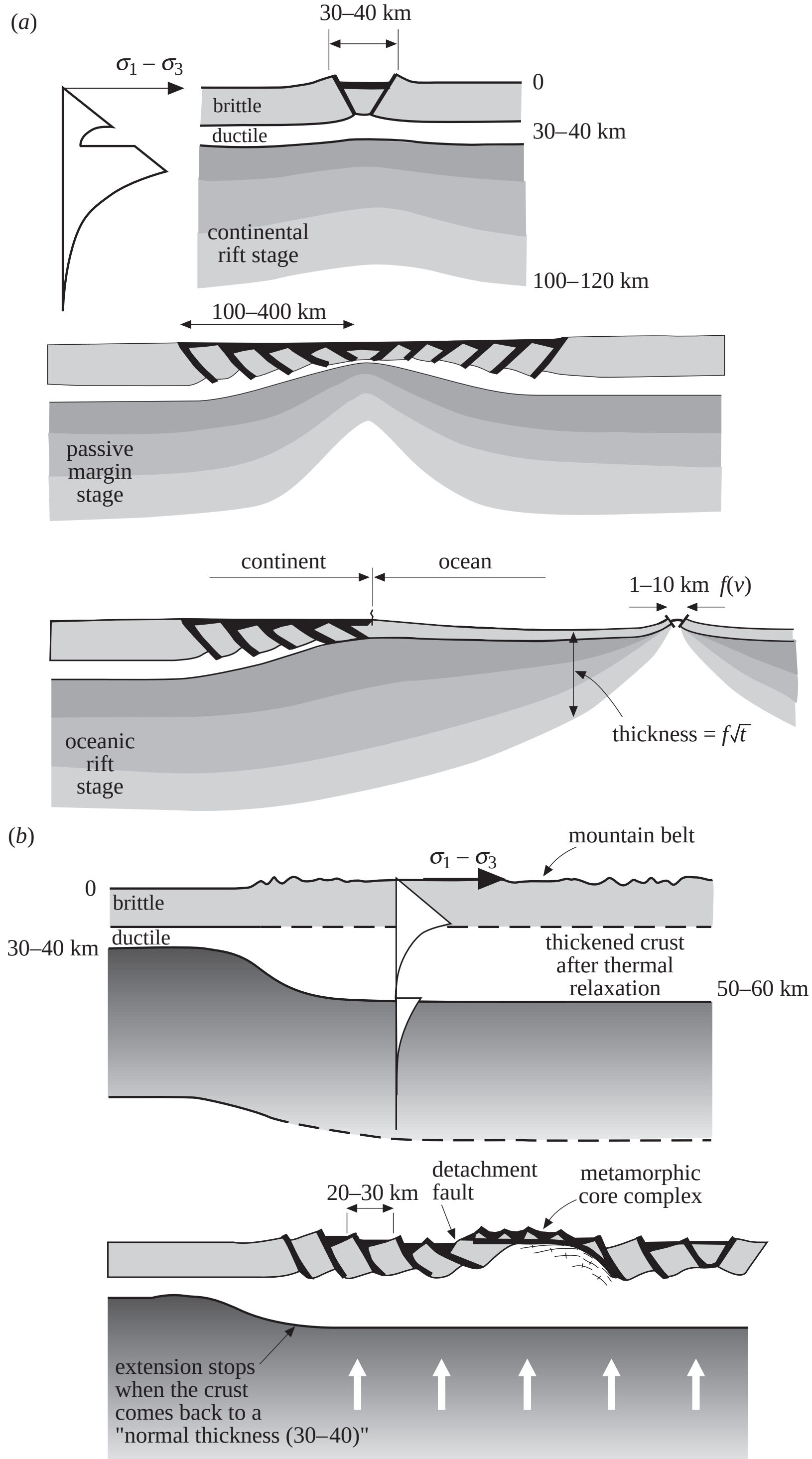

Figure 2. Geological and structural differences between (a) narrow rifts and (b) wide rifts.

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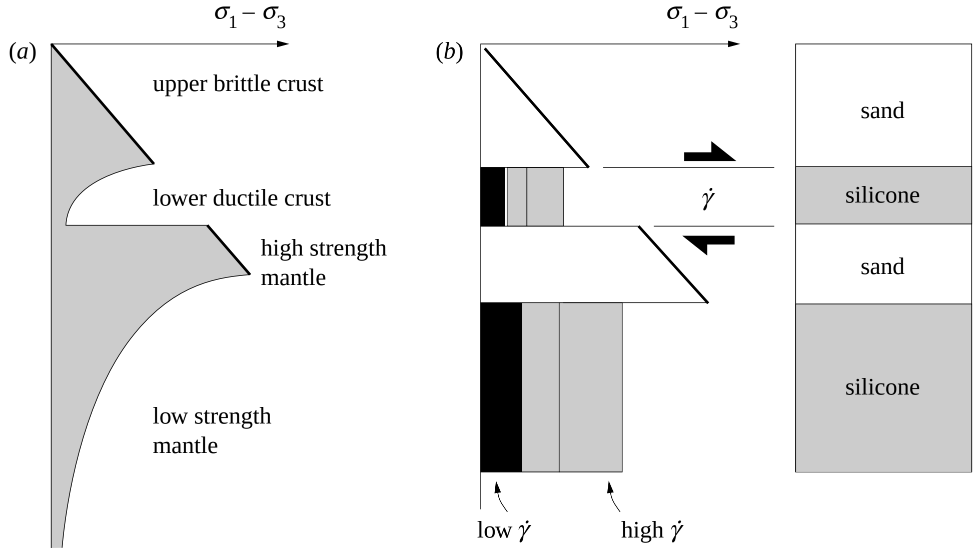

Figure 3. Comparison between strength profiles (a) calculated for a four-layer-type continental lithosphere and (b) obtained in sand–silicone models.

However, a debate remains about some detachment faults that could originate at a shallow angle.

3. Analogue modelling of brittle–ductile systems

For a small-scale model to be representative of a natural example (a prototype), a dynamic similarity in terms of distribution of stresses, rheologies and densities between the model and the prototype is required (Hubbert 1937; Ramberg 1981). The experimental method developed at Geosciences Rennes to study crustal extension started with two-layer brittle–ductile systems made of sand and silicone putty to represent brittle and ductile layers, respectively (Faugère & Brun 1984). Four-layer systems floating above a dense low-viscosity syrup representing the asthenosphere were later introduced (Davy 1986) to simulate deformation of the whole lithosphere. The basic principle of the method consists of simulating simplified strength profiles (figure 3) that incorporate brittle (frictional) and ductile (viscous) rheologies with gravity forces. Scaling relationships between the prototype and the model are obtained by keeping the average strength of the ductile layers correctly scaled with respect to the strength of the brittle layers and the gravity forces.

(a) *Scaling principles*

In the equation of dynamics:

$$\left(\frac{\partial \sigma_{ij}}{\partial X_{ij}}\right) + \rho\left(g - \left(\frac{\mathrm{d}^2\varepsilon_{ij}}{\mathrm{d}t^2}\right)\right) = 0 \qquad (i, j = 1, 2, 3), \tag{3.1}$$

where σ_{ij} are components of stress, ε_{ij} components of deformation, X_{ij} space coordinates, ρ density, g acceleration of gravity and t time. Any modification of the length-scale multiplies the first term by $\sigma^*(L^*)^{-1}$, the second by $\rho^* g^*$, and the third by $\rho^*\varepsilon^*(t^*)^{-2}$, where the exponent $(*)$ refers to model/prototype ratios (e.g. $\sigma^* = \sigma_{\mathrm{m}}/\sigma_{\mathrm{p}}$).

Because equation (3.1) must be respected, we obtain the two following conditions:

$$\sigma^* = \rho^* g^* L^*, \tag{3.2}$$

$$\varepsilon^* = g^*(t^*)^2. \tag{3.3}$$

Inertial forces being negligible in geological processes (Hubbert 1937; Ramberg 1981), only condition (3.2) must be verified. In our experiments, carried out under normal gravity, the gravity ratio is $g^* = 1$. The densities of model materials range between 1100 and 1400 kg m^{-3} and those of rocks between 2300 and 3000 kg m^{-3}. Because model and prototype densities are of the same order of magnitude, the density ratio is $\rho^* \approx 1$. Therefore, condition (3.2) simplifies to

$$\sigma^* \approx L^*. \tag{3.4}$$

In other words, the ratios of stresses and lengths must be nearly equal.

A more detailed analysis, in particular concerning the governing equations, is given by Davy & Cobbold (1991).

(*b*) *Materials*

To represent brittle layers whose behaviour is of Mohr–Coulomb type (Byerlee 1978):

$$\tau = 50 + (\tan 31^\circ)\sigma, \tag{3.5}$$

where τ is the shear stress, 50 the cohesion in MPa, 31° the angle of friction, and σ the normal stress, we use the Fontainebleau sand whose density is $\rho = 1400$ kg m^{-3} and whose angle of friction (ϕ) is in the range 30–33° without significant cohesion. The length-scale ratio being $L^* \approx 10^{-6}$, the cohesion of the sand should be *ca.* 5×10^5 Pa. This value is very small compared to the maximum differential stresses in the models which, for $\tau = 30^\circ$, are given by:

$$\sigma_1 - \sigma_3 = \tfrac{2}{3}\rho g z, \tag{3.6}$$

where σ_1 and σ_3 are the maximum and minimum principal stresses and z is the model depth.

To represent ductile layers, we use silicone putties with variable Newtonian viscosities (from 10^3 to 10^4 Pa s) and densities (from 1100 to 1400 kg m^{-3}). According to the mean strain rate applied to the models, the layers of silicone putty can have extremely variable strengths (figure 3), given by

$$\sigma_1 - \sigma_3 = \mu\dot{\varepsilon} \quad \text{or} \quad \mu\dot{\gamma}, \tag{3.7}$$

where μ is the viscosity and $\dot{\varepsilon}$ the strain rate. The shear strain rate $\dot{\gamma}$ is used as an approximation for experiments in which deformation in ductile layers is nearly simple shear. Therefore, for given relative thicknesses of sand and silicone putty, a wide range of strength profiles can be simulated for different applied strain rates (figure 3).

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(*c*) *Boundary displacements*

In lithosphere experiments, where a two-, three- or four-layer model floats above a low-viscosity syrup (asthenosphere), the basal boundary is free-slip, allowing models to isostatically respond to gravity forces induced by variations in thickness. Displacements are applied at a constant rate along the lateral borders of models (see, for example, Allemand *et al*. 1989; Allemand & Brun 1991; Benes & Davy 1996; Brun & Beslier 1996), or models are allowed to spread under their own weight giving a displacement rate which continuously decreases with time (see, for example, Hatzfeld *et al*. 1997).

In two-layer models designed to represent crustal- or upper-crustal-type sub-systems, displacements are applied over a rigid basal plate at constant rate using moving plastic sheets (see, for example, Allemand & Brun 1991). In spreading-type experiments, nearly free-slip is obtained with a liquid soap film coating the basal boundary (Faugère & Brun 1984; Brun *et al*. 1994).

4. Structural effects of brittle–ductile coupling

(*a*) *Experiments at a constant displacement rate*

Figure 4 presents two series of experiments that illustrate the extreme sensitivity of a two-layer brittle–ductile system to strength variations. In these experiments, displacement is applied at the base of the ductile layer at constant rate along an asymmetric velocity discontinuity.

The three models shown in figure 4*a* have similar thicknesses of brittle and ductile materials with identical rheological properties and the same amount of bulk extension. Strain rates, and therefore ductile strengths, vary by a factor of three (from 0.5×10^{-3} to $1.5\times10^{-3}\ \mathrm{s}^{-1}$). At a low strain rate, deformation remains localized in a single asymmetric graben. At high strain rate, the faulted zone is three times wider. From the bottom to the top of figure 4*a* block tilting increases as a direct function of strain rate.

The three models shown in figure 4*b* have similar thicknesses of ductile material and were run at similar strain rates ($10^{-3}\ \mathrm{s}^{-1}$). Brittle layer thickness, and, therefore, also brittle strength, decrease by a factor of three from bottom to top. At low brittle strength, faulting invades nearly the whole model. At high brittle strength, faulting is localized in a narrower domain. Block tilting increases with brittle strength.

These two series of experiments demonstrate that both brittle and ductile strengths have significant effects on deformation and structural development. An increase in strain rate increases coupling between brittle and ductile layers. As a result, the deformation zone in the brittle layer widens and block tilting is enhanced. In all experiments showing block tilting, the sense of tilting is compatible with the sense of shear in the underlying ductile layer.

The role of gravity is demonstrated in low strain rate models (figure 4*a*, lower case) or low brittle strength models (figure 4*b*, upper case). In these two models, ductile strength is low enough to allow the ductile layer to flow laterally and to compensate for brittle layer lengthening and the associated thinning. However, in the former example of narrow rift type, deformation is localized in a single graben, whereas in the latter example of wide-rift type, deformation is widespread over several grabens. It is noteworthy that, in both models, block tilting is almost negligible.

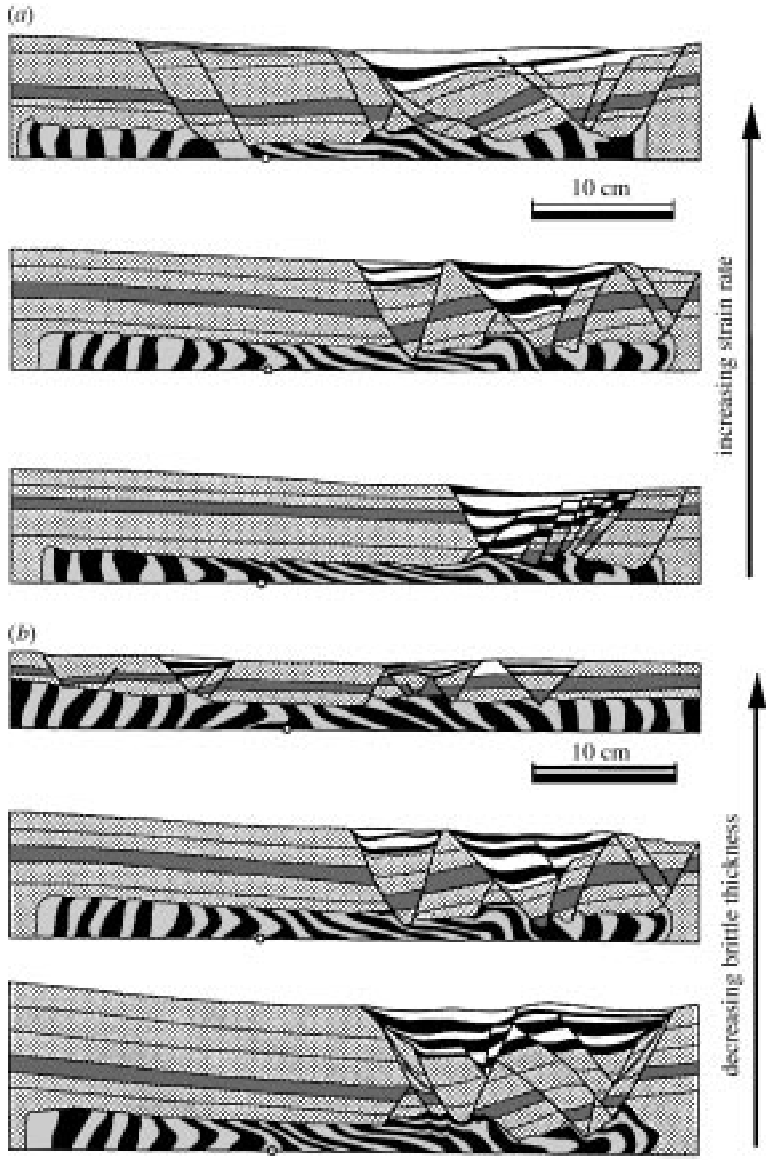

Figure 4. Constant displacement rate deformation on two-layer sand–silicone models showing the variations of extensional structures as a function of (*a*) ductile strength (strain rate) and (*b*) brittle strength (thickness). Modified after Allemand (1990).

(*b*) *Experiments involving gravity spreading*

Models shown in figure 5 correspond to two-layer-type experiments in which the vertical boundary to the right is free to move, allowing models to extend under their own weight. The ductile layer is decoupled from the rigid basal plate by a film of liquid soap allowing free-slip. Spreading rate is controlled by model weight, i.e. driving force, and by silicone viscosity, i.e. resisting force. Spreading rate, therefore, decreases with model thickness during extension. Consequently, strain rate and the

(*a*)

(*b*)

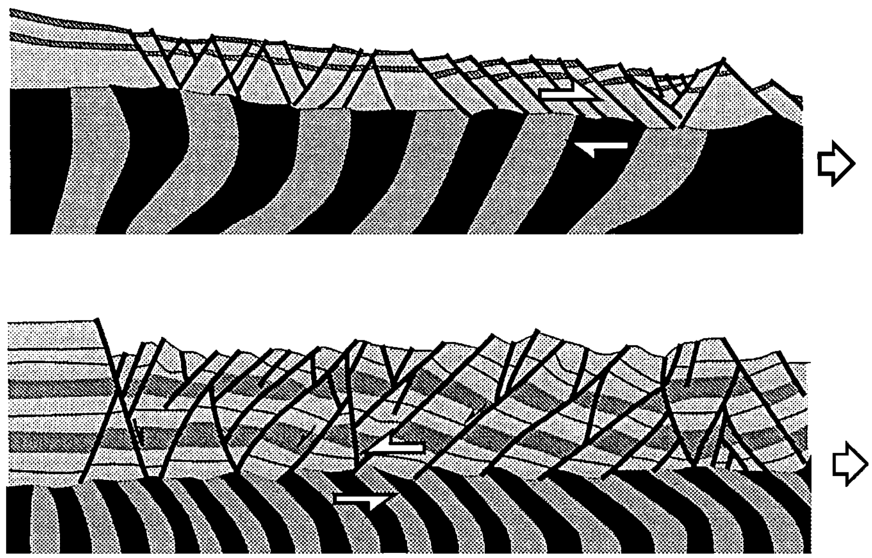

Figure 5. Gravity spreading extension of two-layer sand–silicone models. (*a*) Modified after Brun *et al*. (1994); (*b*) modified after Faugère & Brun (1984).

coupling between the brittle and ductile layers are greatest at the onset of extension. At any time during deformation, strain rate and coupling are at the maximum allowed by the still available driving force.

In the above experiments, faulting invades brittle layers almost entirely and extension is of wide-rift type. The envelope of the brittle–ductile interface remains nearly horizontal during extension. In figure 5*a*, the faulted domain can be divided into two sub-domains: one characterized by conjugate faults and the other by tilted blocks. In figure 5*b*, faulting results dominantly in tilted blocks. The distortion of initially vertical passive markers in ductile layers gives the sense of shear along the brittle–ductile interface (see white arrows), which corresponds to the sense of displacement on faults delimiting tilted blocks. Note that the shear senses are opposite in the two models. The development of tilted blocks and of a wide deformed domain indicates high brittle–ductile coupling at an early stage, in agreement with the models of figure 4*a*.

The upper model of figure 4*b* is also characterized by wide rifting, but with only horsts and grabens instead of tilted blocks. This experiment was carried out at a constant displacement rate far lower than the maximum spreading rate of the models in figure 7. Therefore, in wide rifting mode, the presence of tilted blocks is an indication of high strain rates and of extension dominantly driven by gravity. Conversely, horsts and grabens are indications of strain rates lower than those which would result from pure gravity-driven extension.

5. Necking of a four-layer lithosphere

(*a*) *Continental rift stage*

The Rhine Graben provides a typical example of a narrow rift, whose crustal-scale structure was recently documented by two deep seismic lines shot by the DEKORP

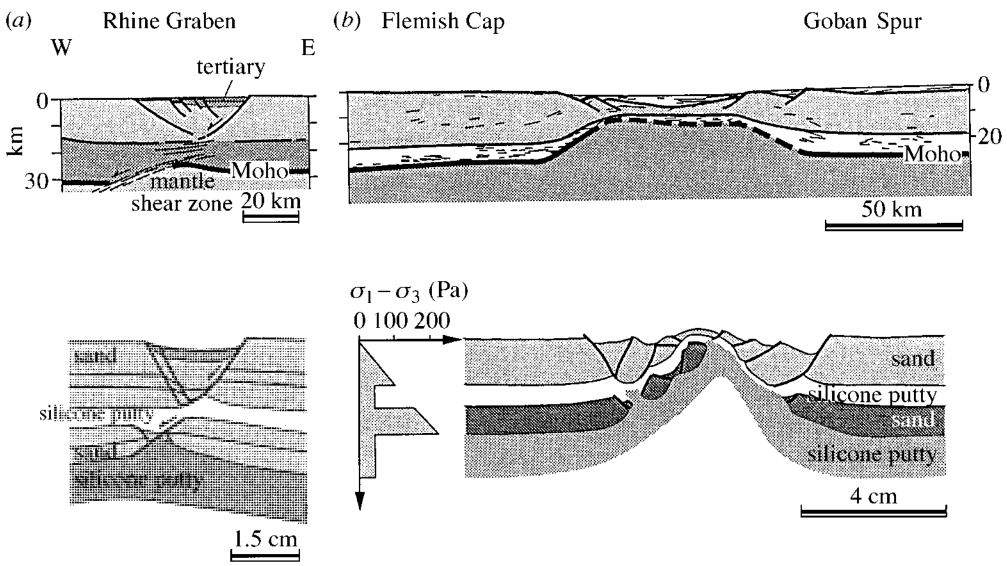

Figure 6. Narrow rifts in nature and experiments. (*a*) Continental rift stage: crustal-scale cross-section of the Rhine Graben modified after Brun *et al.* (1992) and four-layer-type analogue model modified after Beslier (1991). (*b*) Passive margin stage: restored conjugate margins of Canada and Europe modified after Keen *et al.* (1989) and four-layer-type analogue model modified after Brun & Beslier (1996).

and ECORS consortia, in the north and in the south of the Graben (Brun *et al.* 1992). Only the northern section is displayed in figure 6*a* (upper part). Both lines show a strong structural asymmetry with one master normal fault associated with a series of conjugate normal faults. The asymmetry reverses from north, where the master normal fault is located to the east (figure 6*a*), to south, where it is located to the west. The lower reflective crust decreases in thickness below the Graben together with a loss in reflectivity. Below the Graben shoulder, facing opposite to the master fault, gently dipping reflectors cross-cut the Moho with a normal sense of offset, indicating the existence of a narrow mantle shear zone. The upper crustal master normal fault does not appear to be directly connected to the mantle shear zone.

The Rhine Graben structure can be nicely compared with a four-layer-type experiment at low strain rate (figure 6*a*, lower part), which shows a graben whose delimiting normal faults join at the brittle–ductile interface. Such a geometry, commonly observed in experiments, suggests that initial graben width can be used as a measurement of the brittle upper crust thickness (Allemand & Brun 1991). The sand layer, representing a sub-Moho high-strength mantle, displays a region of asymmetrical ductile stretching with a normal-sense offset of the model Moho, below the graben shoulder facing opposite to the master fault. Distributed deformation in the silicone layer representing the lower crust accommodates the two zones of deformation in the surrounding sand layers. In other words, it plays the role of a decollement between brittle upper crust and sub-Moho mantle. Localized mantle deformation is transmitted to the upper crust through this diffuse lower-crust deformation.

The close similarities between model and nature here strongly support the applicability of four-layer-type strength models (figure 1*a*) to narrow rifting, and throw light on the interpretation of the lower crust as a potential zone of decollement. The

Rhine Graben, which has been active since the early Oligocene (40 Ma), is characterized by a small amount of extension, 5 km on average, for an initial rift width of *ca.* 35 km. This gives a low mean strain rate of $1.5 \times 10^{-16}\ \mathrm{s}^{-1}$, which again reinforces the comparison with a low-strain-rate model.

(*b*) *Passive margin stage*

During the last decade, the tectonic interpretation of passive margins has been dominated by the pure shear versus simple shear debate. Pure shear extension is generally advocated through the modelling of subsidence histories (see, for example, McKenzie 1978; Royden & Keen 1980). The apparent symmetry of an entire rift zone, such as the one shown in figure 6*b* (upper part), also favours such an interpretation. Simple shear associated with lithosphere-scale detachment faults cross-cutting the whole lithosphere (Wernicke 1985) is, on the other hand, mainly based on structural arguments and proposed to explain the common asymmetry of fault patterns (see, for example, Lister *et al*. 1986).

Stretching of four-layer-type models of the lithosphere gives broadly symmetrical necking (figure 6*b*, lower case), which corresponds to bulk pure shear. However, structural asymmetries develop internally due to boudinage in the sand layer, representing the sub-Moho high-strength mantle. The two-layer models of figure 4*a* show that an increase in strain rate increases the width of the deformed zone. Similarly, in four-layer models, a single zone of necking of the high-strength mantle layer develops at low strain rate (figure 6*a*) and multiple necking, i.e. boudinage, at higher strain rates (figure 6*b*).

The occurrence of exhumed mantle rocks at the extremity of the Galicia Margin was interpreted in terms of simple shear extension with a detachment fault cross-cutting the whole lithosphere (Boillot *et al*. 1988). Four-layer-type models never show any type of detachment fault, but, conversely, offer a simple explanation for mantle exhumation. As exemplified by figure 6*b*, sub-Moho mantle boudinage evolves in an unstable and asymmetric manner during high-amplitude stretching. A local separation between two boudins of the sub-Moho high-strength layer allows ductile mantle to come into contact with ductile lower crust. Extension becomes localized in this area, leading to extreme stretching and exhumation of ductile mantle.

Several workers had difficulties in accounting for observed subsidence histories using the uniform stretching model of McKenzie (1978). Royden & Keen (1980), Hellinger & Sclater (1983), and many others after them, found it necessary to allow the lithosphere mantle to extend more than the crust. As illustrated by figure 6*b*, sub-Moho mantle boudinage brings a mechanically realistic solution to this problem.

6. Spreading of a thickened lithosphere

(*a*) *Wide rift*

Wide rifting occurs in previously thickened domains where the dominantly ductile behaviour of the lithosphere makes it able to spread under its own weight.

Models presented in § 4 *b* illustrate that two-layer brittle–ductile systems undergoing free-gravity spreading are characterized by strong coupling between brittle and ductile layers, and that consequently they develop domains of tilted blocks (figure 5). This is what is commonly observed in extended regions like the Basin and Range or

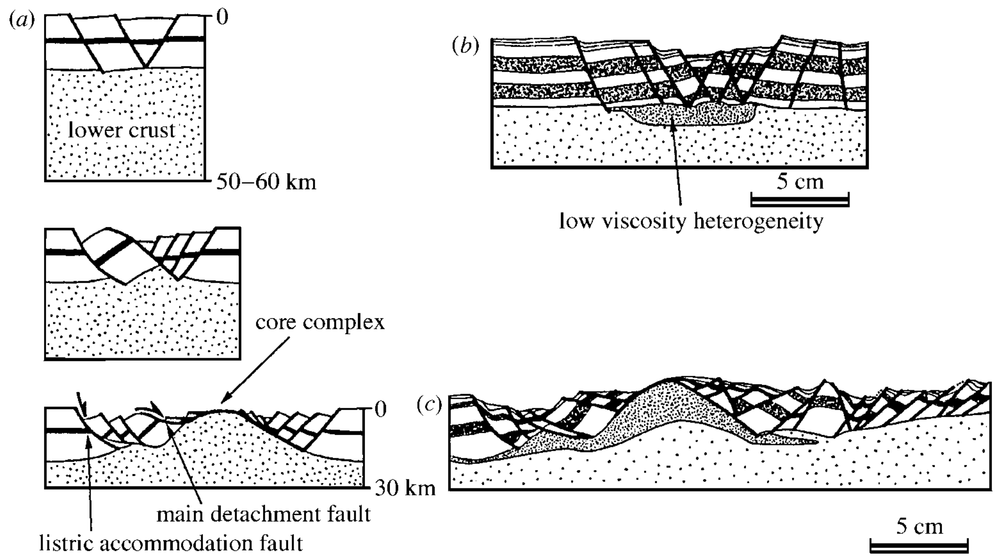

Figure 7. Analogue models of core complexes and detachment faults. (*a*) Conceptual model of core complex development deduced from experiments. Pattern of model extension (*b*) at early stage and (*c*) at core complex stage. Modified after Brun *et al*. (1994).

the Aegean. One of the models presented in § 4 *a* has an initial brittle–ductile thickness ratio of 1.0 (figure 4*b*, upper case), which would allow easy gravity spreading. This model, which was submitted to a rate of extension lower than a free-spreading rate, is characterized by a low brittle–ductile coupling and developed horsts and grabens. Such a pattern of extension is observed in southern Tibet where E–W Tertiary extension resulting from N–S shortening gave birth to a widespread system of N–S trending grabens (Armijo *et al*. 1986).

In the above-quoted experiments, the brittle–ductile interface remains nearly horizontal due to lateral flow of the ductile layer, which maintains constant thinning across the extending zone. This is in agreement with the lack of crustal-scale necking and nearly horizontal Moho in wide rifts.

(*b*) *Core complexes*

Models shown in figure 7 were carried out with the same procedure as those of figure 4, except that a low-viscosity heterogeneity was placed below the brittle–ductile interface in their middle part. The viscosity heterogeneity is only one order of magnitude lower (1.2×10^3 Pa s) than the viscosity of the ductile layer (1.8×10^4 Pa s). In nature, this could correspond to a local thermal anomaly, due for example to a granite pluton or a zone of partial melting.

Figure 7*a* shows the general conceptual model of the development of core complexes resulting from experiments. At the onset of extension (figure 7*b*), a graben-type structure with steep normal faults initiates above the low-viscosity heterogeneity. Then faulting invades the whole model. The stretching rate above the heterogeneity remains higher than elsewhere in the model. At the core complex stage, the amount of horizontal stretching above the low-viscosity heterogeneity is $\lambda_\mathrm{h} = 2.25$, whereas bulk model stretching is $\lambda_\mathrm{h} = 1.00$. Consequently, thinning of the brittle crust becomes stronger in this area, allowing the ductile layer to rise and exhume to form a so-called core complex (figure 7*c*). The core complex is bordered on one side by strongly tilted blocks, and on the other by a detachment fault with a convex-upward shape. Steepening of the footwall is accommodated by one or more listric faults.

These experiments demonstrate that flat-lying detachment faults can result from the rotation of faults that initiate as steep normal faults. Two mechanisms of rotation are involved. The first, confined to within the brittle layer, is similar to the block tilting described in figure 5, where layer-parallel stretching is partly or totally transformed into rigid-body rotation. The second consists of rotation imposed on the faulted layer at its base by the local rise of the ductile layer. Both types of rotation combine to accommodate footwall steepening below the main detachment fault.

These experiments suggest that core complexes and associated detachment faults are not characteristic of a distinct mode of extension, as previously proposed (Buck 1991), but rather correspond to local anomalies in wide rifts. The experiments also demonstrate that in wide rifting, even small lateral variations in ductile lower-crust rheology are able to localize extension.

7. Conclusions

The above review of laboratory extension experiments on brittle–ductile systems leads to the following conclusions.

(*a*) *The modelling technique*

Various types of lithosphere layering can be simulated in the laboratory with simple first-order models made of sand and silicone putty to represent brittle (frictional) and ductile (viscous) rock rheologies, respectively. Experiments strongly support the idea that the main difference between narrow rifts and wide rifts relies on the type of mechanical instability that occurs at a lithospheric scale: i.e. necking versus gravity spreading.

Despite their inherent limitations, sand–silicone models are able to realistically reproduce most, if not all, types of extensional structure known to occur in the upper crust: planar and non-planar faults; grabens; tilted blocks; detachment faults; and core complexes. The sand–silicone technique is, therefore, a powerful tool to study the structural response of an extending brittle–ductile lithosphere to extension.

(*b*) *Narrow rifts*

Narrow rifting corresponds to a necking instability of a four-layer-type lithosphere. The existence of a sub-Moho high-strength layer plays a dominant role on the deformation pattern and history. The initial width of the extending zone is controlled by sub-Moho mantle boudinage. At a low strain rate, i.e. when coupling between brittle and ductile layers is low, a single neck can develop in the sub-Moho mantle, leading to a narrow extending zone at a lithospheric scale. Increasing the strain rate increases the coupling between brittle and ductile layers, allowing multiple necking, i.e. boudinage, of the sub-Moho mantle and consequent widening of the extending zone.

The initial width of individual rifts is, to first-order, a direct function of the thickness of the brittle upper crust, but the upper crust extending zone may contain several rifts, especially when the sub-Moho mantle undergoes multiple necking.

Necking models basically correspond to bulk pure shear at a lithospheric scale. Pure shear is internally accommodated by conjugate zones of simple shear, which develop s along brittle–ductile interfaces. These shear zones cannot be considered as simple shear detachment faults cross-cutting the whole lithosphere. The separation of boudins in the sub-Moho mantle naturally leads to ductile mantle exhumation without the need for a detachment fault initially cutting the whole lithosphere.

(*c*) *Wide rifts*

Wide rifting results from gravity-spreading-type instabilities of a previously thickened and dominantly ductile lithosphere, where the highest strength layer is the brittle upper crust. In models, the maximum rate of spreading depends on the driving force, i.e. model weight, and the resisting force, i.e. viscosity of ductile layers. Rates of extension significantly lower than the maximum rate of spreading favour the development of horsts and grabens. Conversely, rates of extension approaching the maximum rate of spreading favour the development of tilted blocks. The vergence of tilted blocks is controlled by the sense of shear along the brittle–ductile transition.

(*d*) *Core complexes*

Core complexes do not correspond to a particular mode of extension but rather to local anomalies within wide rifts. Models sustain the idea that they probably develop in zones of the upper crust located above heterogeneities of the ductile lower crust that are weak enough to localize stretching. Consequent local stronger thinning of the brittle upper crust is compensated by the uprise and exhumation of the ductile lower crust. Low-angle detachment faults that accommodate the exhumation of core complexes in models result from the sequential development and progressive flattening of initially steeply dipping normal faults.

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Discussion

R. Buck (*Lamont-Doherty Earth Observatory, USA*). I was curious about the core complex mode that was shown in one of the models. Did the model require a heterogeneity, or simply a very weak viscous layer?

J.-P. Brun. We tested various types of perturbation, in terms of boundary conditions and rheology, in a set of more than 40 experiments. We found that a low-viscosity heterogeneity seated below the brittle–ductile interface is the only way of producing a core complex-type structure in two-layer brittle–ductile models. In models with a simple, very weak layer, the brittle–ductile interface remains horizontal and no core complex develops.

K. McClay (*Royal Holloway College, University of London, UK*). I would like to support an important point about variations in strain rate. We see, both in the analogue models and in the computer models, that what we are really imposing is a boundary condition of either constant velocity or constant displacement. The analogue models adjust internally to variations in strain rate and heterogeneities. The point to consider for the computer models is whether they can do that as well. Can they continue building those heterogeneities, because, as Professor Brun has shown, heterogeneities focus high-strain-rate deformation and the initiation of rifting? Most rifts develop on old fold belts with internal structures, such as the Red Sea, and the Gulf of Suez that developed along the old Nile shear system. I would also like to add a comment about wide rifts and the development of core complexes; some people have recently described core complexes on the Eritrean margin, but there is not really a wide rift in the Red Sea.

J.-P. Brun. Many people try to put core complexes in many places, but I know core complexes only in the Basin and Range and in the Aegean and similar areas of greater age, such as the Variscan Belt or the Caledonides. Core complexes are always in regions that were previously mountain belts. However, Jeff Karson (this issue) suggests that we have core complexes in the ocean.

N. Kusznir (*University of Liverpool, UK*). I would like to ask Professor Brun a question which is probably more of a statement. When he showed the conceptual model of silicone–sand–silicone, he made the point that the stresses in the silicone, if it is strained fast, can exceed those in the sand. That, of course, in a sense illustrates the limitation of his analogue model because presumably in the Earth what would happen is that the material exposed to a big stress would actually fail by brittle deformation, so he would need a silicone sand or something.

J.-P. Brun. Yes! This was illustrated by the series of experiments applied to salt tectonics in the North Sea. At a high strain rate, faults cut through the silicone layer representing the salt.

K. E. Louden (*Dalhousie University, Halifax, Nova Scotia, Canada*). Professor Brun's models contain a large degree of complexity in the faulting process, which is difficult to include in the theoretical computer models. On the other hand, the computer models include temperature effects from thinning the asthenosphere, which are difficult to include in his models. Is there any possibility that we can at some point look forward to a model that includes both of these types of effects?

J.-P. Brun. In choosing certain thicknesses of brittle and ductile materials we make implicit assumptions about the initial thermal state. But because sand cannot change into silicone, and conversely silicone cannot change into sand, models are unable to take into account rheological modifications due to temperature variations during deformation. This is the greatest limitation of analogue models. The numerical models of rifting, presented by Roger Buck (this issue), had well-localized shear bands. These are not yet faults, but I am confident that in the future, with new computers and new codes, numerical models will become able to really simulate faulting. We will then stop doing experiments with sand and silicone putty and give them back to children.

N. Kusznir. Professor Brun made the point that the lower crust under the Rhine Graben is granulite, but it appears to be formed not by brittle deformation but by plastic deformation. Would he like to argue in support of that?

J.-P. Brun. The geometrical analysis of the Rhine Graben seismic lines suggests that deformation, in the lower crust below the Graben, is rather diffuse and therefore ductile. This is in contrast to the underlying mantle where a series of low-dipping reflectors connected to an offset of the Moho suggests the existence of a shear zone. The rheological behaviour would therefore be of the 'localizing' type if not brittle. On the other hand, we know, from the Vosges and Black Forest Massifs, that granulites are present in the Variscan crust. This is more an apparent contradiction than a provocation. It could simply mean that granulites at a very low strain rate—here $1.5 \times 10^{-16}\ \mathrm{s}^{-1}$—could behave in a ductile manner.

D. McKenzie (*Bullard Laboratories, University of Cambridge, UK*). If the lower crust is warm, it will certainly flow. The question is: is its viscosity higher or lower than that of the mantle beneath the Moho? In somewhere like the Rhine Graben it seems quite likely that actually the lower crust is weaker than the upper mantle. The fact that the behaviour can be reproduced with honey, which I suspect under Professor Brun's conditions is weaker than the silicone putty, only illustrates that more forcibly. So that really isn't the difficulty. There and in the Basin and Range we have similar flow in the lower crust. But such flow could occur even if the lower crust was stronger, rather than weaker, than the upper mantle.

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N. Kusznir. My recollection of rheological strengths is that while both feldspar and anorthosite are much stronger than quartz, they are still substantially weaker than olivine. The lower crust should therefore be weaker than the mantle even if lower crustal rheology is not dominated by quartz.

D. McKenzie. I have a slight worry. I don't remember the experiments accurately enough to know what the anorthite number of the anorthite was, but the value is important. A good rule of thumb is that the melting temperature is actually what controls the rheology. The albite proportion of feldspar strongly reduces its solidus temperature, and hence is likely to weaken the material. It is likely that the anorthite number of the lower crust is high, and therefore that it is strong. From an experimental point of view, it is important that the creep experiments are carried out on material that has a known mineral composition.

J.-P. Brun. It is also important to consider the strain rate argument. Extension in the Rhine Graben started 40 million years ago and the total amount of extension is around 5 km, assuming an initial Graben width of 30–35 km. From these values the horizontal stretching rate is around 10^{-16} per second. At such a low strain rate even granulitic materials could be ductile.

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