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INTRODUCTION—PROCESSES OF CONTINENTAL RIFTING

PAUL MORGAN and BRIAN H. BAKER

Lunar and Planetary Institute, 3303 NASA Road One, Houston, TX 77058 (U.S.A.)
Center for Volcanology, University of Oregon, Eugene, OR 97403 (U.S.A.)

This volume is the outcome of a conference on the Processes of Planetary Rifting which was convened for three days in December 1981 (Anonymous, 1981). The purpose of this conference was to bring together theoretical modellers, geologists, geochemists and geophysicists to give the modellers direct exposure to the constraints imposed by some of the available data. Similarly, workers primarily involved in data collection were exposed to the implications of models to data interpretation and speculation. We attempted to advance the understanding of the processes of rifting by encouraging a multidisciplinary approach to the problem.

As a focus for discussion, participants were encouraged to present their ideas in the context of two end-member processes for rifting, active rifting, in which rifting is a result of a thermal upwelling of the asthenosphere, and passive rifting, in which rifting is a passive response to a regional stress field (Sengör and Burke, 1978; Baker and Morgan, 1981). We concentrated on continental rifts because they provide an accessible evolutionary record of rifting possibly diagnostic of the processes of their formation.

No attempt has been made to make this volume complete in the sense of covering all aspects of rifting, or indeed all ideas presented at the conference. We believe, however, that this incomplete mixture of models, data and speculation provides a representative sample of current thinking on continental rifts. In this introduction we present a commentary on the models of the processes of continental rifting, and some data relevant to these models, with special reference to papers presented in this volume. We conclude with suggestions for future multidisciplinary studies.

MODELS OF RIFTING

To understand the processes and causes of continental rifting it is necessary to examine the consequences of the modification of the lithosphere during rifting. Universal features of rift zones are crustal (lithospheric) extension with graben formation, and anomalous crust and upper mantle, usually interpreted as the result of lithospheric thinning or asthenospheric diapirism. In a broad sense it is possible to estimate the magnitude of deviatoric stresses required to form a graben, but it is

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much more difficult to constrain the cause and effect relationships of processes in the lower lithosphere/asthenosphere system.

Bott (1981) and Bott and Mithen (this volume) have examined the development of rift grabens, without considering how they were initiated, by assuming that the subsidence of a graben is related to the loss of gravitational energy as a wedge of brittle upper crust descends, accompanied by the outflow of ductile material in the lower crust. This mechanism requires a substantial deviatoric extensional stress, but less separation of the rift sides than is required if the graben is assumed to subside in response to necking by uniform lithospheric stretching (e.g., see McKenzie, 1978). For continental rifts, Bott and Mithen (op. cit.) calculate that a deviatoric stress on the order of 100–200 MPa (1–2 kbar) is required. During rift development, therefore, this stress must be developed within the upper crust either as an intraplate stress related to plate boundary forces, or as a result of lateral density contrasts within the plate. Anomalous mantle structure and domal uplifts are common features of rifts, so we first consider the latter mechanism for stress generation.

An extensional stress field is generated by a surface load provided by uplifted topography together with the corresponding upthrust caused by the anomalous low density mantle which isostatically supports the elevated topography, as was recognised by Bott (1971) and Artyuskhov (1973). The magnitude of this deviatoric stress in the upper crust has been calculated to be on the order of 200 MPa (2 kbar) (Kusnir and Bott, 1977; Bott and Kusnir, 1979; Artyushkov, 1981; Bott, 1981; and others). Further calculations of this stress are given by Crough (this volume), Neugebauer (this volume), and Turcotte and Emerson (this volume), the general conclusions of which are that lateral density contrasts associated with anomalous lithospheric structure can cause significant deviatoric stresses in the upper crust capable of initiating or propagating a graben. Crough (op. cit.) also considers a general compressive stress field due to ridge push on the plate margins, and concludes that the gravitational body forces developed by uplift in oceanic swells is never great enough to overcome the ridge-related compressive stresses, with the result that oceanic swells never rift. However, the gravitational body forces developed in continental swells uplifted 1 km or more are sufficient to overcome a regional compressive stress field, and rifting can occur. As we emphasise later, however, the genesis of each rift should be examined with reference to its specific stress field, and how these stresses change with time.

Before discussing the effects of regional stresses we examine the mechanisms proposed to generate lithospheric thinning and uplift. Three basic mechanisms have been proposed: (1) thermal thinning in which the lithosphere is static, but material is removed from the base of the lithosphere by heating, conversion to asthenosphere, and removal in an asthenosphere convection system; (2) mechanical thinning, or lithospheric stretching in which lithospheric material moves laterally in response to a regional extensional stress field, and the asthenosphere rises passively to fill the void created by the thinning lithosphere; and (3) asthenospheric diapirism in which the asthenosphere penetrates the lithosphere driven by the gravitational instability of the less dense asthenosphere under the more dense mantle lithosphere, and flow occurs in both the lithosphere and the asthenosphere. The third mechanism is essentially a response to a perturbation in the lithosphere/asthenosphere system, possibly initiated by either thermal or mechanical thinning, but driven primarily by gravitational energy.

Thermal thinning of the lithosphere, or effective thinning by heating the lithosphere to cause uplift, can occur by simple conduction from a sub-lithospheric heat source, heating of the lithosphere by penetrative convection of magma into the lithosphere, or convective heating of the base of the lithosphere over the rising limb of an asthenosphere convection system. Simple conduction is discounted as a primary mechanism because it is too slow to produce observed uplift rates (Mareschal, this volume). Magmatic heating is thought unlikely to be the primary mechanism because of the large volumes of magma required, and because the heat would be concentrated in a narrow zone (Mareschal, op. cit.; Turcotte and Emerman, this volume). The efficiency of convective thinning depends on the magnitude of the anomalous heat flux supplied by the asthenospheric convection system. Both Spohn and Schubert (this volume) and Wendlandt and Morgan (1982) have shown that the lithosphere can be thinned to crustal levels in a few tens of millions of years with an increased heat flux across the base of the lithosphere of 5–10 times the background heat flow. From steady state models of mantle plumes (Parmentier et al., 1975; Yuen and Schubert, 1976) this increase is reasonable. Turcotte and Emerman (op. cit.), however, using stagnation point theory, conclude that convective thinning would be too slow to be a primary thinning mechanism. Bailey (this volume) discusses the role of mantle degassing in the thermal and magmatic evolution of a rift, and this process may assist, or locally dominate over thermal thinning. The problem with mantle degassing is that the dominant varieties of igneous rocks found in rifts are also found in other tectonic environments. Thus, although it seems inevitable that thermal thinning must occur over asthenosphere hot spots, there is no agreement at present on the efficiency of this process, the resulting rate of thinning, and of the role of diapirism.

Models of simple geometrical stretching of the continental lithosphere, based on the concepts of McKenzie (1978) invariably predict downwarp (Blackwell and Chockalingam, 1981), and while these models may have some application to the subsidence of rift floors, in their present form they cannot explain uplift associated with rifting. Simple geometrical stretching, in which the amount of stretching is constrained to be constant with depth without regard to changing rheology, is unrealistic, however, and models with more complex rheologies have yet to be tested. Even where simple stretching models have been applied to areas of subsidence in extensional settings, the models have been modified by effectively increasing the stretching in the lower lithosphere relative to the crustal stretching (equivalent to an additional heat input into the lower lithosphere) to explain features of these areas (e.g., Royden and Keen, 1980; Sclater et al., 1980; Chénet and Montadert, 1981). Some uplift can be produced in the unthinned lithosphere by lateral heat flow from the thinned portion of the lithosphere, which could produce a narrow uplift on rift margins (Jarvis, 1981), but would not explain the broad swells commonly associated with rifts.

The mechanical thinning models do not yet consider the response of the asthenosphere during its passive upwelling beneath the stretched lithosphere (decompression melting, diapirism, etc.), and must be regarded as incomplete. Realistic modelling of the consequences of lithospheric stretching remains to be performed.

An alternative approach to studying the feasibility of passive stretching of the lithosphere by a regional stress field is to consider how a suitable stress field can be generated. Neugebauer (this volume) calculated the stress field produced by an indenter during collision orogeny such as that represented by the Alpine-Himalayan system and its possible effects on extension in the Rhine and Baikal zones. He concluded that the stresses were not sufficient to cause rifting, but might serve as a supporting boundary condition. Bott (this volume) proposes the trench suction forces (see Forsyth and Uyeda, 1975) on either side of Pangea in the early Mesozoic as a possible mechanism capable of generating a rifting stress field. In an earlier paper Bott (1982) estimates that these forces may be sufficient to allow graben formation in regions where the elastic lithosphere has been thinned over a mantle hot spot. Such a model requires restrictive tectonic conditions for rifting, and cannot be applied to many Cenozoic rifts. In view of the uncertainties in the forces on the plates (e.g., see Forsyth and Uyeda, op. cit.) and in the strength of the lithosphere, it is probably premature to make any general conclusions about the role of regional stresses and passive stretching in rifting. Rifting *requires* extension, however, and this extension must be permitted by the regional stress field.

The final rifting mechanism that has been proposed is the development of an asthenospheric diapir penetrating the lithosphere, an extreme example of which is represented by lithospheric delamination (Bird, 1979). The static physical configuration is obvious: more dense mantle lithosphere overlies less dense asthenosphere. The development of a diapir is strongly dependent on the effective viscosities in the lithosphere/asthenosphere system, and these viscosities control the rate of diapir growth. Numerical calculations by Bridwell and Potzick (1981), Neugebauer (this volume) and Mareschal (this volume), show that significant asthenosphere diapirism should occur in a few millions to a few tens of millions of years if the effective viscosity of the lower lithosphere is reduced to be on the order of 10^{20}–10^{22} Pa s (10^{21}–10^{23} poise). Thus for diapirism to be effective, thinning to initiate diapirism, and thermal preconditioning of the lithosphere by heating or stretching are prerequisites of the diapiric thinning.

These basic models of rifting provide insight into the possible processes of rifting and predict evolutionary histories of rifting by which the mechanisms can be tested. Additional models are often required to link the possible processes to observations (e.g., see DeRito et al., this volume; Morgan, P., this volume; Morgan, W.J., this volume; Sheridan, this volume). The rifting process needs to be considered in two parts: (1) What is the initiating process—active or passive? (2) Once the lithosphere has begun to thin, what are the roles of magmatism, diapirism, and of the tectonic environment? The models inevitably require refinement before they can have any general application, but they can be used to indicate the kinds of data that will be useful to constrain rifting processes. Conversely, geological, geochemical and geophysical models based on data can provide direction for the improvement of the theoretical models of rifting processes.

CONSTRAINTS ON THE MODELS

A test of a model is that it should fit any geological, geochemical and geophysical data that are available: many of the papers in this volume provide some of the constraints for the theoretical models, but it is obvious that in most cases the work was not designed to obtain those constraints.

The majority of geological papers are general descriptions of rift features, for example of Proterozoic rifts of the Canadian shield (Easton, this volume), the midcontinent region of North America (Keller et al., this volume; Green, this volume; Gilbert, this volume), the Oslo graben (Russell and Smythe, this volume; Schönwandt and Petersen, this volume), the Rio Grande (Golombek et al., this volume), and West Africa (Fitton, this volume). Discussion of these "failed" rifts shows the considerable variation in structure, volcanism, and geophysical signatures, which result from the influence of pre-existing lithosphere structure and contemporaneous plate dynamics.

The evolutionary history of older rifts is often not well established owing to difficulties of dating and overprinting by subsequent structural events, and little can be deduced about the mechanism by which they were formed. Keller et al. (this volume) suggest that the Keweenawan event was related to Grenville orogenesis, and that Eocambrian and Mesozoic rifts of North America are "failed arms" associated with continental break-up of a Precambrian and Pangean continent respectively. They point out that many major sedimentary basins are located on rifts, although DeRito et al. (this volume) note that the timing of subsidence cannot commonly be explained by thermal subsidence after rifting, and present a model of reactivation of uncompensated mass excess, the result of crustal densification during rifting, to form the basins. For presently active rifts, Golombek et al. (this volume) deduced a passive (lithospheric extension) process for the development of the Espanola basin of the Rio Grande rift, whereas Logatchev et al. (this volume) conclude from their comparison of the Kenya and Baikal rifts that both were formed by active (plume-type) mechanisms.

Several papers cover questions concerning structures and their mechanisms. Mohr (this volume) examines the validity of the Morton-Black model for antithetic fault swarms, and concludes that data from the margins of Afar depressions do not support it. This work highlights the fact that little work has been done on the mechanics of normal fault swarms and of the block tilting found in them. There is an inadequate basis for downward projection of the surface structure of rifts; somewhere there must be a rift with the deeper parts exposed by erosion. Wood (this volume) reports evidence that rift zones can jump laterally, and notes that the jump distances are approximately equal to the lithosphere thickness, suggesting that jumped rifts are passive rifts, controlled by lithosphere properties rather than by the asthenosphere. The role of pressurised cracks in causing small-scale rift structures is examined by Pollard et al. (this volume), with the conclusion that these features are predicted well by elastic theory. Unfortunately, it is difficult to extrapolate these results to large scale continental rifts.

The basic styles and causes of rifting are examined in three papers: Bailey (this volume) advocates mantle degassing; Sheridan (this volume) suggests that mantle convection is pulsed by heat transfer out of the core giving periods of rifting and/or rapid sea-floor spreading; Milanovsky (this volume) reviews the changing character of rifts through geological time.

Browne and Fairhead (this volume) and Bermingham et al. (this volume) propose largely on the basis of gravity data that the Ngaoundere rift (Cameroon "line") extends across Central Africa into the Sudan and links with isolated domal uplifts (e.g., Darfur) to form a Central African rift system which is in an early stage of development. They interpret the scarcity of domal uplifts and major faults to indicate that the rifting is passive. In this contribution, and that of Girder (this volume), the authors are influenced by the existence of zones of negative Bouguer gravity anomalies that they interpret as incipient rift zones. Crough (this volume) uses similar data to constrain the geometry of the deep lithospheric structure associated with continental rifts and swells, concluding that broad asthenospheric upwarps are indicated by the data rather than narrow dyke-like upwarps.

Heat flow and temperature are fundamental parameters either implicitly or explicitly used in all rift models, but as illustrated by a review of heat flow data in Cenozoic rift systems by P. Morgan (this volume), these parameters are not well defined by the available data. The main problem with heat flow studies in rifts is the effect of heat transfer by hydrothermal circulation, which makes the near surface thermal field very complex, an example of which is given by Crane and O'Connell (this volume). Conductive electrical anomalies are commonly taken as evidence for high crustal temperatures in rift zones, but as discussed by Jiracek et al. (this volume), there can be many sources of the conductivity anomalies. An electrical anomaly in the Rio Grande rift does not correlate with a lower crustal magma chamber in the rift (postulated to exist primarily from seismic data). Jiracek et al. (op. cit.) interpret the anomalous conductivity to be associated with a ductile zone at the base of the brittle upper crust. Thus, although high electrical conductivities in the crust may indicate elevated temperatures, the technique is poorly calibrated at present. Uplift is viewed as a thermal parameter by P. Morgan (this volume) and W.J. Morgan (this volume). W.J. Morgan concludes that uplifts can persist for up to 100 m.y. in rocks resistant to erosion by isostatic uplift in response to erosion, which may explain many contemporary shield exposures devoid of significant sediments. Many of these attempts to define the thermal parameters in the lithosphere show promise but the available data do not rigorously constrain the models.

The most rigorous constraints on lithospheric structure in rift zones come from seismic data, and Olsen (this volume) reviews the available data from Cenozoic rifts. Thinned crust and anomalously low compressional wave velocities in the upper mantle, generally interpreted as evidence for asthenospheric upwelling, are indicated by the seismic data in the Cenozoic rifts. At present the most detailed data are available from the Rhinegraben, which indicate a complex crust–mantle transition zone. More data are required, expecially from the classic Kenya rift to understand the details of crust and lithospheric structure associated with rifting. Detailed refraction results from the northern Mississippi Embayment, which is generally regarded to be a Late Precambrian rift, reactivated in the Mesozoic, are reported by Mooney et al. (this volume) which indicate that the crust was not permanently thinned during rifting, but that extensional thinning was compensated possibly by lower crustal magmatism, resulting in a normal thickness crust with a high seismic velocity (and by implication density) in its lowest levels. A similar seismic structure with actual crustal thickening has been recently reported for the rift-like Snake River Plain (Smith, 1982, et seq.). These results illustrate the need for inclusion of effects of magmatic processes in the theoretical models.

RESOURCES ASSOCIATED WITH RIFTING

In recent years it has been realised that a distinctive suite of economic minerals is associated with rifts. These include coal, lignite, and hydrocarbons, lead–zinc, baryte, fluorite, and molybdenum deposits, and evaporites of sodium and potash. Their occurrence is due to the formation of anoxic, saline lakes, aided by volcanism and hydrothermal activity, and to plutonism at depth. The interaction of physical, chemical, and biological processes in the rift lacustrine environment is reviewed by Robbins (this volume), and the relationship of porphyry-molybdenum deposits to granitic plutons of Rapakivi type in the Oslo rift is described by Schönwandt and Petersen (this volume). A complete model of the processes of rifting must include the chemical and physical conditions to allow these resources to develop.

The geothermal resources of the Salton trough and the Rio Grande rift are compared by Swanberg (this volume), who concludes that high temperature resources are likely to be found in mature rifts with silicic volcanism, rather than in younger rifts with only basaltic volcanism. Forced convection in rift basins can give rise to low temperature waters that issue above sills and constrictions between interconnected basins in the rift system. Rifts are favourable for the formation of accessible hydrothermal systems because of the often intense igneous activity, tl localisation of hydrothermal circulation by faults, and the presence of shallo plutons. With respect to the processes of rifting, geothermal systems can account fo a significant portion of the heat loss in a rift zone (e.g., see Crane and O'Connel this volume; Morgan, P., this volume).

CONCLUDING REMARKS

Because rifting gives rise to such a variety of geological processes, the study o rifts provides excellent opportunities for relating deformation, igneous activity, an sedimentation to their underlying tectonic causes, and to understand the interactio of the asthenosphere with the lithosphere. It is unlikely that any one of the model suggested in this volume or elsewhere will prove to be the sole mechanism of rifting and though in any one rift one mechanism may be the primary cause of rifting, th interplay of subsidiary mechanisms may confuse simplistic interpretations of avail- able data with respect to cause of rifting. It is probable that thermal thinning and/o diapirism can cause the extensional stress required for rifting; however, rifting wil not occur unless the regional tectonic regime will permit the sides of the rift to diverge. On the other hand, passive plate extension could cause rifting in isolation, but the extension and rifting is likely to be localised where the lithosphere is weakest over an existing thermal anomaly. Where asthenospheric diapirism occurs, which is essentially a response to thinning of the lithosphere by thermal thinning or plate extension, the effects of diapirism may completely mask the initiating mechanism.

Anomalous heat transfer into the lithosphere, diapirism and magmatism must all play linked roles in rifting, together with a deviatoric stress field that will permit extension in a developing rift. The models are useful in that they permit idealised processes to be quantified and tested, but it is clear that better knowledge of lithosphere properties is needed, particularly of mantle viscosity and its temperature dependence.

Now that a variety of models of rifting is available, it is necessary to identify the critical data that would permit discrimination between them. These data include better information of deformation rates, the duration, volume, and character of igneous activity, and the previous history of the lithosphere and its contemporary dynamics. Quantifying the evolution of rifts can be done more easily and completely in still active rifts in which the record is accessible ~~and can be dated with precision~~, and in which the anomalous lithosphere structure survives. By juxtaposing theoretical models and geological, geochemical and geophysical descriptions in this volume we hope that the dialogue between geologists and theoretical modellers will be enhanced; that theoreticians will constrain their models better using geological data, and that geologists will search for data that provide better constraints.

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