Papers Presented to the

CONFERENCE ON THE PROCESSES OF PLANETARY RIFTING

CHRISTIAN BROTHERS' RETREAT HOUSE

NAPA VALLEY, CALIFORNIA

3-5 DECEMBER, 1981

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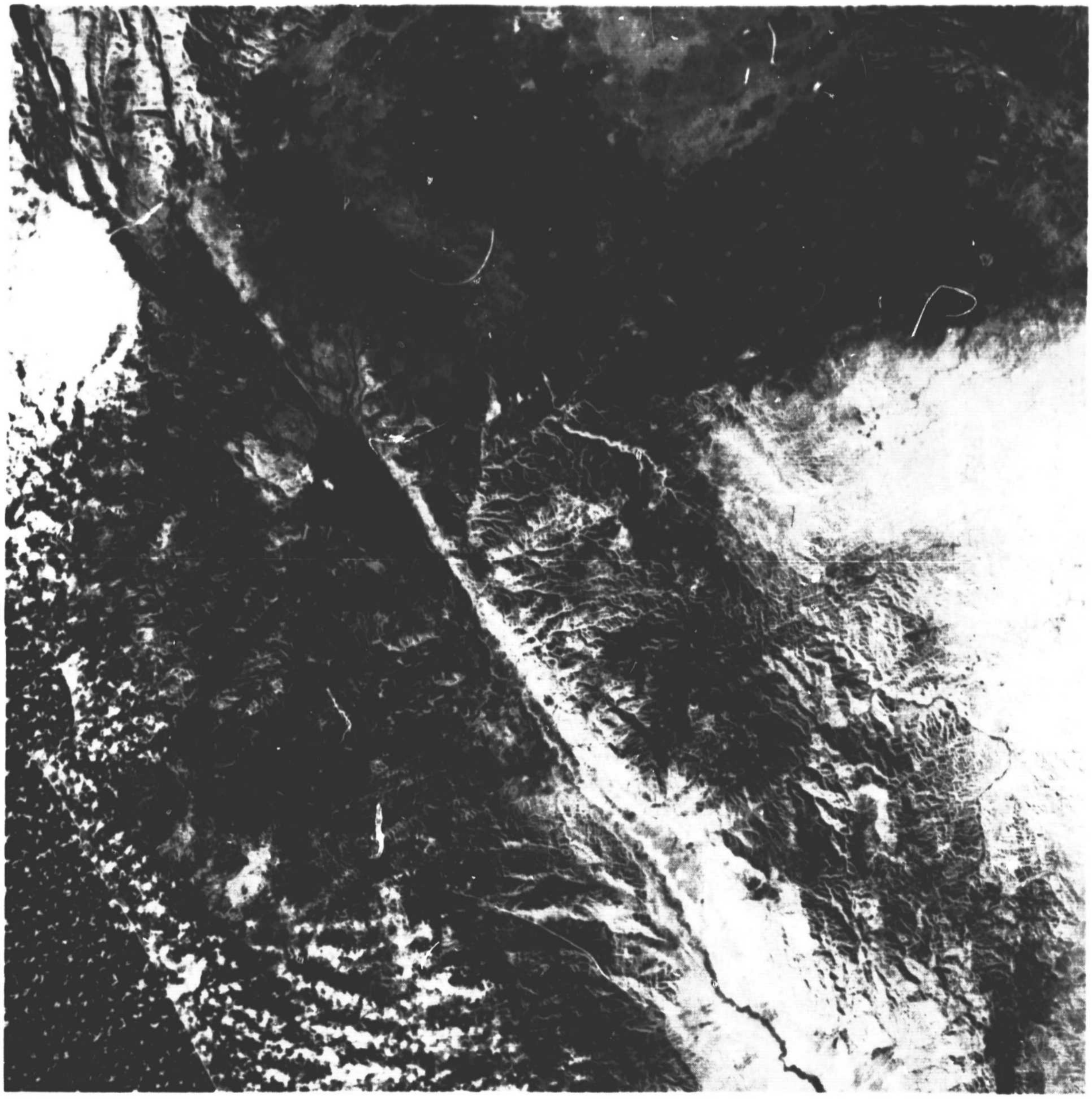

Skylab camera photograph of the northern portion of the Dead Sea rift in the Sea of Galilee region. Top of photograph is approximately northeast. Each side of photograph is approximately 160 km of ground coverage. Circles northeast of (above) the Sea of Galilee are volcanoes of the Golan Heights. (Skylab photo frame SL3-46-209.)

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PREFACE

The last two decades have seen a great increase in the volume of geological, geochemical and geophysical data from continental rift zones, but these new data have brought little agreement among earth scientists on the basic processes of rifting. The Conference on the Processes of Planetary Rifting has been convened to review and assess the current state of our knowledge of continental rifting. The title for the conference was chosen to expand the discussion to other planets where rift-like features have been observed, but the basic thermo-mechanical properties of the planet are different from the Earth. The format of the conference is unorthodox in starting with the speculation and models of rifting processes, followed by the data presentations. This format was chosen to stimulate discussion on the constraints placed by the data on the models, and to indicate areas where new data are required. A session on resources was included to emphasize the economic importance of rift zones, and to provide additional data to constrain the models. Papers were solicited which address the basic problems of the processes of planetary rifting, and these have been organized into the following sessions:

I. Speculation as the Origins and Development of Rifts

II. Rifts on Other Planets

III. Tectonics

IV. Geology

V. Chemistry of the Lithosphere

VI. Physics of the Lithosphere

VII. Resources Associated with Rifting

VIII. Our State of Ignorance and its Remedy (Debate, no talks; abstracts submitted in this session are for publication only.)

Papers are reproduced in this volume in the order of the preliminary session program.

The active members of the program committee for the conference have been: Brian H. Baker, Co-convenor (*University of Oregon*), Scott Baldridge (*Los Alamos National Laboratory*), Tom Crough (*Purdue University*), Tom Giordano (*New Mexico State University*), Randy Keller (*University of Texas at El Paso*), Ivo Lucchita (*U.S. Geological Survey, Flagstaff*), Michael Mayhew (*Business and Technology Systems, Inc.*), Paul Morgan, Co-convenor (*Lunar and Planetary Institute*), Ric Wendlandt (*Lunar and Planetary Institute*), and Chuck Wood (*NASA, Johnson Space Center*).

Logistic and administrative support for this conference has been provided by P. H. Jones (*Projects Administrator, Lunar and Planetary Institute*). This volume was compiled by K. Hrametz (*Technical Editor, Lunar and Planetary Institute*).

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HOTSPOT TRACKS AND THE EARLY RIFTING OF THE ATLANTIC

W. Jason Morgan, Dept. Geological and Geophysical Sciences, Princeton Univ., Princeton NJ 08544

Many hotspot tracks appear to become the locus of later rifting, as though the heat of the hotspot weakens the lithosphere and tens of millions of years later the continents are split along these weakened lines. Examples are the west coast of Greenland - east coast of Labrador (Madeira hotspot), the south coast of Mexico - north coast of Honduras (Guyana hotspot), and the south coast of West Africa - north coast of Brasil (St. Helena hotspot). A modern day analog of a possible future rift is the Snake River Plain, where the North American continent is being "pre-weakened" by the Yellowstone hotspot track.

This conclusion is based on reconstructions of the motions of the continents as shown in Figure 1. The relative motions of the plates were determined from magnetic anomaly isochrons in the oceans (and from guesses where there are no identified anomalies, e.g., the Gulf of Mexico). The motion of one plate was chosen ad hoc to best fit the motions of the plates over the hotspots, but once the motion of this one plate was chosen, the motions of all the other plates were prescribed by the relative motion constraints. It is assumed that all of the hotspots shown in Figure 1 are active for the entire 180 My, and predicted tracks based on the plate motions over the hotspots are shown with tick marks marking every 30 My. The dashed lines are the plate boundaries active at each time, and the figure shows how tracks switch plates as boundaries migrate over hotspots (e.g., Verde at 120 My and Meteor at 60 My). The hotspots are permitted to wander slightly (at about 3 mm/yr) relative to the average "fixed frame" in order to better fit presumed hotspot features. For example, note that the Fernando hotspot has a position at 1 °S in the 180 My reconstruction (at that time in Louisiana) and is at 4 °S in the present reconstruction (at Fernando de Noronha off the coast of Brazil); the sequence of dots in the first frame of the figure shows how much Fernando has wandered during this 180 My. The other clusters of dots in the first frame show how much each of the hotspots must wander in order to adequately fit presumed tracks.

A complete presentation of the method used above and a discussion of the comparison of the predicted tracks with observations are given in Morgan (1981) for the Atlantic and Indian oceans. The rotations used to construct Figure 1 are from Morgan (1980), which are slight revisions to the earlier written 1981 reference. The conclusions of this study are: (1) Hotspots are reasonably fixed and form a convenient refernce frame; wander of individual hotspots is only about 3 mm/yr relative to the average frame. (2) Hotspots appear to persist for 100 to 200 My (or

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perhaps more); some die away while others appear to increase in intensity during this interval. (3) Many of the continental breakups appear to be along lines "pre-weakened" by a hotspot trace. In addition to the cases given above, the Gulf of Aden (Comoros hotspot), the east coast of Madagascar - west coast of India (Crozet hotspot), and the Naturalist Plateau - Gaussberg Plateau (Kerguelen hotspot) are examples from the Indian Ocean. Also, rifting on continents which does not persist to become a new ocean can also result from hotspot weakening. For example, the Sirte Basin in Libya is at the 120 My standstill position of the Cameroon track (F. B. Van Houten, personal commun., 1981).

Hotspot tracks can affect the study of rifts in yet another way. The passive margins of the Atlantic are interpreted as examples of "mature" rifts, however some of the variability of the margins is due to the effects of later hotspot crossings of the passive margins. For example, the Cape Fear Arch in North Carolina is where the (weak) Bermuda hotspot crossed the margin 60 My ago. Some differences in the margin at Cape Fear as compared to the Georgia Embayment and Baltimore Canyon area should be interpreted as due to this later heating and uplift caused by the hotspot and not as differences in the initial rifting process. Another example is the hotspot crossing in the Georges Bank area, marked by the prominent New England Seamounts. Note that by coincidence, two hotspots cross this same area -- the Verde hotspot about 160 My ago and the Meteor hotspot about 110 My ago. (These two hotspot crossings account for two of the three groupings of ages in the dates of the White Mountain Magma Series, but not the 210 My grouping.) The uplift and erosion of the continent caused by hotspot tracks here is discussed by Crough (1981), and the sediment from this erosion event should be an important factor in the nearby continental margin histories.

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Morgan, W.J.

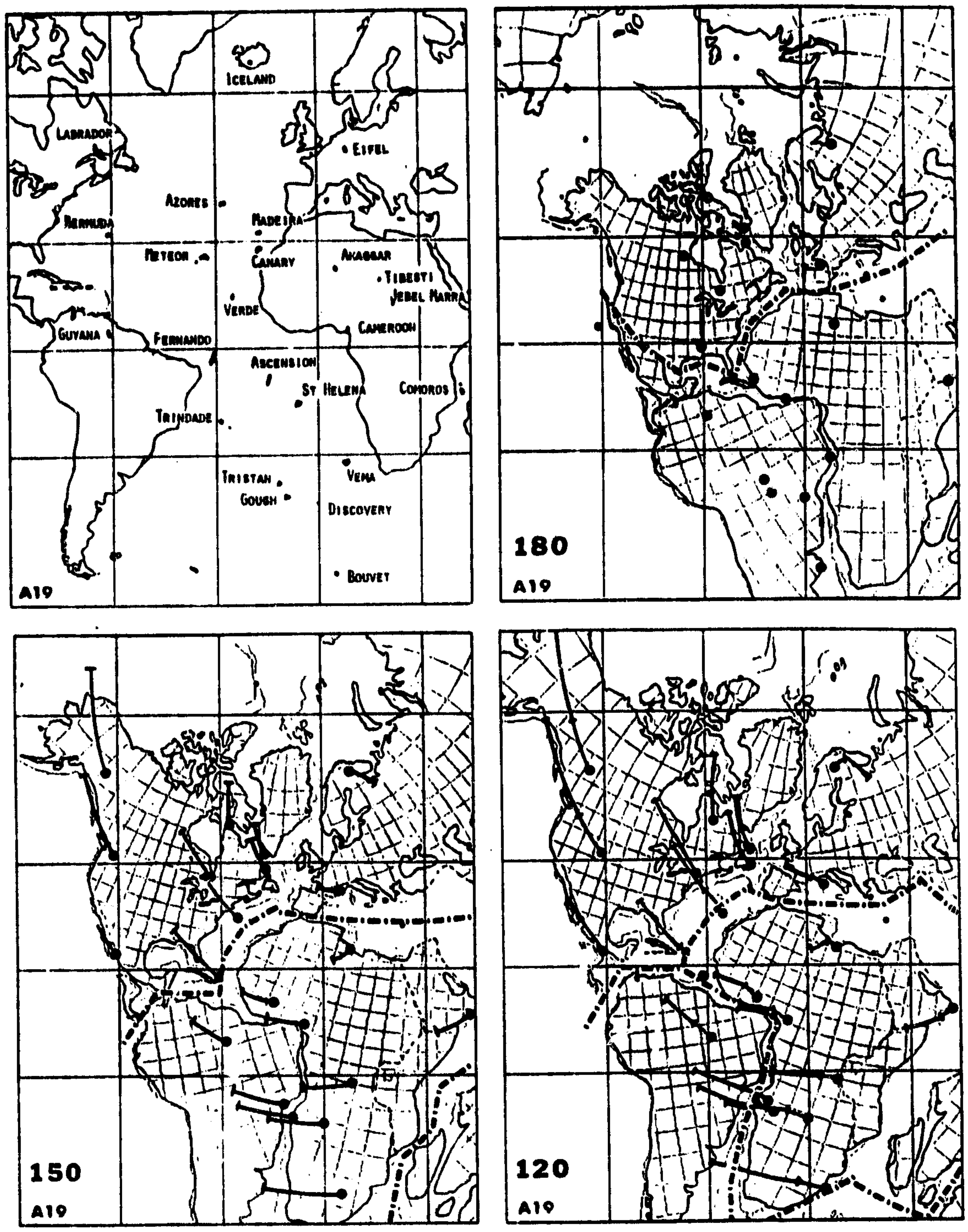

Figure 1. Reconstructions of the plates around the Atlantic from 180 My to present, based on rotations from Morgan (1980).

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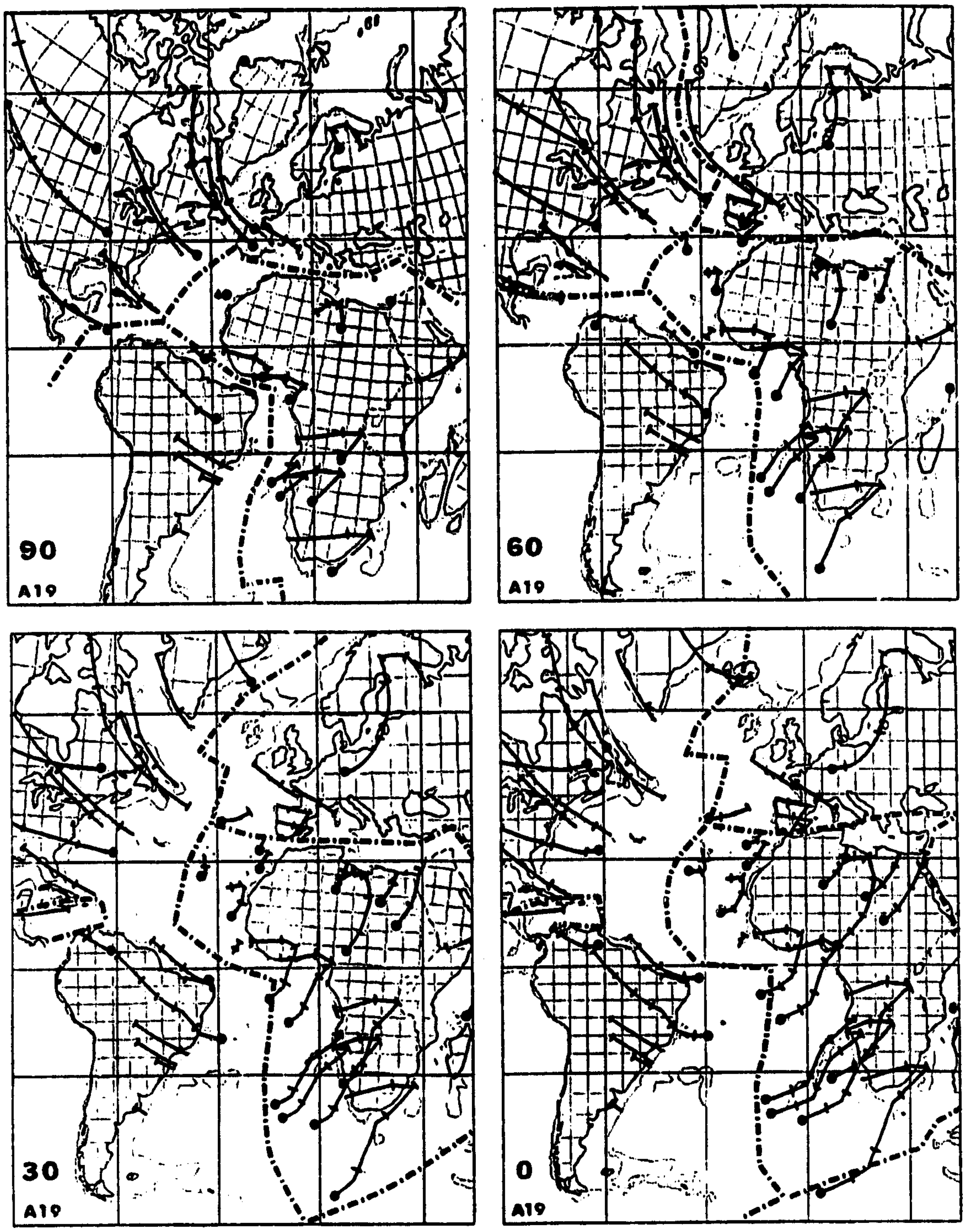

The plates move in a fixed hotspot frame. Hotspot tracks are heavy lines with tick marks every 30 My. Mercator projection.

RIFTS - TENSIONAL FAILURES OF THE LITHOSPHERE. D.L. Turcotte, Department of Geological Sciences, Cornell University, Ithaca, N.Y. 14853

The median valleys of the world-wide ocean ridge system represent fully developed planetary rifts. Many aspects of these rifts are reasonably well understood. They represent the locus of sea-floor spreading and passively tap the upper mantle rock beneath. Ocean ridges are not associated with the ascending limbs of mantle convection cells. Mantle rock passively flows upward to fill the gap created by sea floor spreading. Anomalous sections of the ridge system such as Iceland, the Azores, and the Galapagos islands may reflect regions of active mantle upwelling. Active mantle upwelling may also be associated with oceanic islands and some intracontinental volcanism.

In many if not all cases, oceanic rifts originate as continental rifts. The Red Sea is an example of the transition. There are also examples of continental rifts which do not reach the oceanic stage. These fossil rifts (or graben valleys) have a wide spectrum of ages. There are also a number of examples along the margins of the Atlantic where three rifts form a triple junction; two branches open to form the Atlantic and one branch (the failed arm) becomes a fossil rift or aulacogen (Burke 1976, 1977).

An important question regarding rifts is their cause. There are essentially two hypotheses. The first is that rifts are caused by mantle plumes (Morgan 1972) and the second is that rifts are caused by failure of the lithosphere under tensional stresses (Turcotte and Oxburgh 1973). In essentially all cases volcanism is associated with rifting. This volcanism is probably due to pressure release melting associated with the vertical ascent of mantle rock. In many cases crustal domes are associated with rifts. Examples are the Ethiopian and East African Swells on the East African Rift system. The near circular plan form of crustal domes makes it attractive to associate these features with axisymmetric mantle plumes. Tensional tectonics is also associated with rifting. The graben-like structure of rifts is the result of tensional strain. Fossil rifts exhibit a residual extension of about 10-30 km. Seismic focal mechanisms indicate tensional stresses in rifts. The essential question concerns cause and effect. Are the tensional strains the result of mantle plumes or are the volcanics the result of a tensional failure of the lithosphere?

Let us first consider the plume hypothesis. The primary evidence favoring nearly fixed mantle plumes is the linear age progression of the Hawaiian Emperor island and seamount chain. This progression can be easily explained if the Pacific plate is moving over a fixed mantle hot spot. Although there is no direct evidence for a mantle plume beneath Hawaii, indirect evidence comes from the anmalous geochemistry of the basaltic rocks. Normal ocean ridge basalts are believed to originiate from a well-stirred upper mantle reservoir (Jacobson and Wasserburg 1979). It has recently been suggested (Hofmann and White 1981) that the anomalous chemistry is due to the recycling of imperfectly mixed subducted oceanic lithosphere. The subducted lithosphere may lie on the compositional boundary at a depth of 650 km. The heat flux from the lower mantle heats the subducted oceanic crust until it becomes buoyantly unstable and forms an upper mantle plume. Subducted sediments and basalts altered by sea water cause the observed anomalous chemistry (enriched light rare earth elements, etc).

Other oceanic islands tend to exhibit linear age trends (i.e. the Pitcairn-Tuamotu chain) but these are generally not as well defined as the Hawaiian-Emperor chain (Turcotte and Oxburgh 1978). Some fraction of, but probably not all, oceanic islands are likely to be associated with upper mantle plumes.

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Since continental volcanic rocks are generally altered by interaction with the continental crust it is difficult to ascribe continental volcanic rocks to different classes. It has not been possible to determine whether the complex chemistry of the volcanic rocks associated with continental rifts is due to an anomalous mantle source or is due to the melting and assimilation of continental crustal rocks. It would not be surprising, however, if some fraction of continental volcanic rocks can be attributed to upper mantle plumes. The essential question is what fraction?

We next turn to the alternative hypothesis for continental rifts; that they are the result of the tensional failure of the continental lithosphere. There are many sources of stress in the continental lithosphere. It is quite clear that the upper part of the thermal lithosphere (the elastic lithosphere) acts as a stress guide on geological time scales (Caldwell and Turcotte 1979, Watts et al. 1980). The plate boundary forces that drive plate tectonics (ridge push, trench pull) contribute to intraplate forces (Solomon et al. 1975, 1980). Thermal stresses can be important as well as forces due to erosion and sedimentation (Turcotte and Oxburgh 1976). Because the earth is an ellipsoid membrane stresses are generated in plates that change latitude after they have been formed (Turcotte 1974). It has been suggested that these membrane stresses cause continental rifting (Oxburgh and Turcotte 1974, Freeth 1979, 1980). As plates evolve the geometrical incompatibility of the plate boundaries requires intraplate stresses and deformation (Dewey 1975). It was suggested by Atwater (1970) that this type of incompatibility of plate motions is responsible for the broad zone of deformation in the western United States. If intraplate stresses cause continental rifting then the associated volcanism is a secondary process. The volcanism is associated with horizontal extension in much the same way that the basaltic volcanism at normal mid-ocean ridges is due to sea floor spreading.

In order to decide between the two alternative hypothesis for continental rifts it is necessary to study various examples. In the western United States the Rio Grande Rift is the eastern boundary for a broad zone of crustal deformation. The Snake River plain may also be a continental rift but it is covered by extensive volcanics. Suppe et al. (1975) suggested that the tectonics of the western United States could be explained in terms of two plumes, the Yellowstone plume and the Raton plume. However, it is impossible to explain all the volcanics of the western United States in terms of two localized plumes. The extensive volcanics in the Basin and Range province appears to be directly associated with the broad zone of crustal extension in this area. Similarly the Rio Grande Rift appears to accomodate crustal extension rather than being associated with a mantle plume.

Other rifts that appear to have a direct tectonic affinity are the Rhine graben and the Lake Baikal rifts. The Rhine graben appears to be a direct result of the stress field caused by the continental collision between the African and Eurasian plates that caused the development of the Alps (Illies and Greiner 1978, 1979).

The Lake Baikal rift appears to be associated with the broad zone of deformation caused by the continental collision between the Indian and Eurasian plates.

The case for Africa is less clear. The East African Rift does not have a clear relationship with plate boundary forces. However, it can be argued that "ridge push" forces associated with the Red Sea and the Gulf of Aden may be responsible for a tensional failure of the African plate.

It appears that the weight of observational evidence favors the hypothesis

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that intraplate tensional stresses cause continental rifting and that the associated volcanism is a secondary process. It follows that rifts on the moon and Mars are also extensional features associated with tensional stresses in the elastic lithosphere of these bodies.

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Continental Rifting: An Active or Passive Mechanism?

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Continental rifts have a number of characteristic features in common like the confined but elongated extensional tectonics accompanied by vertical movements, volcanism and a thinned crust and lithosphere. While the sedimentary record of rift zones reveal the sequence of relative vertical movements, the igneous character and age of volcanic rocks provide importand information on the deep seated processes in rift zones. The setting of continental rifts in the frame of subduction, collision and orogeny points to a close interaction of rifting and plate dynamics.

In order to study the mechanism of rifting a number of numerical models on alternative processes have been applied and discussed in the view of the above boundary conditions.

Considering the process of lithospheric thinning as the diapiric development and uprise of less dense material the optimum viscosity-depth distribution and upper mantle structure was investigated (Woidt & Neugebauer, 1981). The parameters in question were the period of initiation of the optimum wavelength and the wavelength itself. It is demonstrated on the base of a five layer linear model that adequate wavelengths require a high viscous crustal layer while the period of initiation leads to appropriate time spans only for lower lithosphere viscosities smaller than 10^{23} poise.

Following this line the uprise of density instabilities can be predicted in shape and rate by special finite element technique (Woidt & Neugebauer, 1980). For a sequence of diapirs and corresponding gravitational compensation the related stress and flow fields will be presented. These models show diapiric uprise to be a very efficient model of lithospheric thinning and induced local tensional stresses in the top layers (Neugebauer & Ochmann, 1981).

The vertical and lateral development of crustal failure in response to uplift will be presented by means of three dimensional numerical models with elastic plastic rheology (Neugebauer & Temme, 1981). The results lead to the conclusion that rifts are not likely to be lateral propagating cracks but rather develop associated with vertical movements. This is basically a result of the nonelastic properties of subcrustal material.

After this sequence of models on the consequences of diapiric uprise attention becomes attracted by the question where the above mechanism could be started. The linear multi-layered viscous models of the lithosphere-asthenosphere system reveal information on the increase of the instability amplitudes for each boundary layer. Thus triggering of the density instability by initial disturbances on different boundaries can be investigated. It turns out, that structural heterogeneities at the crustal layer might cause the development of diapiric instability at the top of the asthenospheric layer. However, an instability originated at the low density

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layer itself exhibit the faster development.

Beside the influence of structural heterogeneities at the crust on rifting the possible consequences of continent-continent collision for the rifting mechanism have been modelled on the base of finite element plate models with nonlinear rheology (Neugebauer et. al. 1981). The stress field in the plate in response to indenter like loads is discussed as a function of the amount and shape of the indenter. Under certain conditions it is possible to include a local extensional stress regime in the plate. The dominant stress regime indicates strike-slip conditions in the plane. The zone of low tension is confined and situated distant from the collision zone, the stress level reached a rather low value already. The influence of collision orogeny on continental rifting might be very low and not likely to exceed the effect of triggering by structural changes below the taphrogenic level.

This aspects are discussed by means of the tectonic seeting and development of continental rift structures in general and for the Rhine Rift System in particular.

In the light of rift development and the numerical experiments it is suggested that continental rifting in the sence of taphrogenesis and lithospheric split up requires the mechanism of diapiric uprising material. Passive splitting in the sence of lateral crack propagation through the lithosphere is very unlikely.

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MECHANISMS OF GRABEN FORMATION AND SOURCES OF CAUSITIVE STRESS, M.H.P. Bott and D.P. Mithen, University of Durham, Durham DH1 3LE, U.K.

Terrestrial grabens form in response to horizontal deviatoric tension in the upper brittle part of the lithosphere. Subsidence occurs by normal faulting and flexure, with some compensating uplift. The faulting may or may not be listric, but many rift valleys and fault-controlled sedimentary basins appear to be bounded by plane normal faults between which a downward narrowing wedge subsides. Graben formation implies that the brittle faulted layer passes down into ductile material which takes up the fault motion at depth by flow. Grabens tend to form in regions of relatively high heat flow where the brittle layer is probably at most 20 km thick.

The subsidence is driven by the overall release of gravitational energy as a downward narrowing wedge subsides between complementary rim uplifts or as the fault blocks subside and rotate. Elastic strain energy is actually increased by normal faulting wherever the crust is under overall compression as occurs below a few kilometres depth. Energy is also dissipated by friction on the faults and by underlying flow. The energy budget during wedge subsidence places constraints on the amount of subsidence possible under given conditions. Simple calculations show that subsidence would be completely inhibited if the coefficient of sliding friction on the fault plane is about unity, but significant subsidence can occur for a coefficient of about 0.1, which is realistic for fault gouge. Friction on the faults may be further reduced if pore pressure of water counteracts the normal stress.

Energy budget calculations show that the maximum possible amount of subsidence is greater for narrower grabens and for higher applied deviatoric tension. Sediment loading increases the amount of subsidence by a factor of two to three. Assuming that fault friction is negligible, it can be shown that sediment-filled graben can subside by up to 5 km for a 20 km width and 100 MPa (1 kbar) tension, or for a 40 km width and 200 MPa tension. The graben width may be determined either by bending stresses on the downthrow side of the initial fault or by pre-existing lines of basement weakness. The above calculations are based on a simple theoretical model of a brittle layer above a fluid substratum. Finite element modelling in general supports the above conclusions but suggests that there may be additional complications for more realistic underlying rheologies.

A substantial and persistent deviatoric tension in the uppermost brittle part of the crust is a prerequisite for terrestrial graben formation. Bending stresses and membrane stresses, which can probably be relieved by transient creep, appear to be inadequate. The most obvious source of persistent stress arises from the plate boundary forces. The forces at ocean ridges cause compression, but those at convergent plate boundaries cause tension, with the slabpull force affecting the subducting plate and trench suction affecting the overriding plate. These forces are to some extent counteracted by local

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resistance to plate motion. A problem arises because the deviatoric stresses caused by plate boundary forces are at most about 20 to 30 MPa acting across the thickness of the lithosphere, which seems to be too small to produce graben which subside several kilometres. However, if deviatoric stress in the lower lithosphere is relieved by slow creep, as seems probable, the stress becomes concentrated into the uppermost elastic part of the lithosphere where it may exceed 100 MPa. Such stress concentration into the uppermost elastic layer is most effective in regions of high heat flow where the elastic layer is thinnest, and is least effective in shield regions where the layer is thickest. In this way, deviatoric tension adequate to form observed grabens can originate under suitable palaeogeographical conditions from the boundary forces at convergent margins.

To give examples, the slabpull force may have given rise to the Carboniferous fault-controlled basins of Britain, which formed on the northern shelf of a narrow closing ocean. Under the rather special palaeogeographical setting when subduction has occurred on opposite sides of a large continental plate, such as Pangaea during the Permo-Triassic, the whole continental plate would be expected to be in tension as a result of trench suction acting on opposite sides without much local resistance. This situation may account for the prevalence of graben formation in the early Mesozoic, culminating in the continental break-up of Pangaea. It should be emphasized that tension stemming from plate boundary forces can only occur on planets where the mechanism of heat escape from the deep interior involves plate motions.

Tension in the crust can also develop in uparched plateau uplift regions which are isostatically supported by an underlying low density region. A horizontal deviatoric tension results from the combined effect of surface load and upthrust of the low density compensating region. This situation occurs in regions such as East Africa as a result of mantle hot spot activity causing thinning and thermal expansion of the lithosphere. The deviatoric tension produced by this mechanism is greatest where the upper elastic part of the lithosphere is thinnest. Stresses of at least 200 MPa can develop, which is adequate for graben formation. It is suggested that such tension associated with elevated East Africa has given rise to the rift valleys of that region. This mechanism is independent of plate boundary forces and it may be applicable to planets with an immobile lithosphere.

AN EXTENSIONAL MODEL OF SUBMARINE RIFTING

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A model for the formation of mid-ocean ridges is presented in which the central rift zone is interpreted as a belt of localized extension of width 2W centred on the ridge axis. Upwelling of the asthenosphere occurs passively below the zone of extension at the rate required to conserve mass. At distances greater than W from the ridge axis the lithospheric plates move at a uniform horizontal velocity u_o. The model is thermally and mechanically consistent and predicts the surface heat flow and topography across the rift zone plus the vertical thermal structure and the thickness of the lithosphere below the zone of extension. Measurements of ridge heat flow and topography can be used to constrain estimates of the width of the zone of intrusion.

There have been few developments of thermal models of the oceanic lithosphere since the "plate model" was introduced by McKenzie (1967). The continued success of this model, in accounting for large scale oceanic bathymetry and heat flow variations, is due to the fact that far from the ridge crest the heat flow and topography are not very sensitive to details of the initial conditions. Indeed the prediction of all subsequent thermal models of the lithosphere converge with those of McKenzie's plate model for sea-floor ages greater than about 10Ma (and less than ~80Ma). Consequently a physical understanding of the mechanism of plate formation is only possible by examining features of the ocean floor close to the ridge axis - say for ocean floor ages less than 20Ma.

Close to the ridge axis the plate model becomes untenable: predictions of heat flow are infinite at the ridge axis while at sea-floor ages of 4 to 5 Ma they are as much as 30% too low. In otherwords the central anomaly is too narrow and too large. This is because in the plate model, upwelling mantle is injected along the axial plane only - and at infinite velocity. Since the extensional model allows upwelling over a broad region at finite velocities, it predicts both finite values for the axial heat flow and a broader central anomaly. As a result, predicted heat flow values can account for the observed high values at ages of 5Ma or more.

The mathematical model is composed of two zones: the inner zone ($|x| \leq W$, where x is the horizontal coordinate normal to the ridge axis) in which the lithosphere is stretched; and an outer zone ($|x| \geq W$) in which the plates move uniformly with a constant velocity u_o. The upper surface is maintained at a constant temperature T = 0°C, while the lower surface is held at $T = T_1$ the assumed constant temperature of the asthenosphere below. Within the inner zone a pure shear, constant strain rate, velocity field is imposed in which the horizontal component, u, varies linearly from u = 0 at the ridge axis to $u = u_o$ at $x = W$. Similarly the vertical component of velocity, v, decreases linearly from $V = u_o$ at a depth, a, to v = o at the upper surface. For simplicity V_o is assumed constant for all $x \leq W$ and the horizontal diffusion of heat is neglected.

Within the inner zone the steady vertical temperature distribution is

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obtained analytically and is found to depend on the one dimensionless parameter.

$$G^1 = aV_o/\kappa \tag{1}$$

where a, the vertical extent of the model, also represents the lithospheric thickness at old ages, and κ is the mean thermal diffusivity of the lithosphere. Conservation of mass requires that

$$V_oW = u_oa \tag{2}$$

so that for a given u_o and a (generally known quantities) V_o is inversely proportional to W. Using Equation (2), G^1 may be re-expressed as

$$G^1 = \frac{a^2u_o}{\kappa W} \tag{3}$$

and hence, assuming κ is known, the one free parameter in the problem is W the half-width of the intrusion zone. By varying G^1 to obtain the best agreement between model predictions and observations, W can be estimated from Equation (3).

Measured values of heat flow close to ridge crests show a wide scatter, which is generally attributed to hydrothermal circulation in the crustal rocks of the young sea floor. Since this circulation acts to lower the conductive heat transport, even reliable heat flow measurements on ridge flanks must represent a lower bound to the total heat flow. In fact the highest measured values may prove to be the most reliable. In order to produce a heat flow one standard deviation above Sclater et al's (1976) reliable means, at ages of 3 and 5 Ma, a model with $2W \approx 80$ km is required, while a heat flow just marginally higher than these reliable means requires $2W \simeq 40$km. These two models predict axial heat flows of 13HFU and 18HFU respectively.

The topography of the ridge flanks, as inferred from such models, follows approximately the same linear dependence on the square root of sea-floor age as do the ocean floors of the major ocean basins. However the model ridge topography is flat across the top of the intrusion zone. This feature is a consequence of the assumption that the intrusion velocity V_o does not vary with x and the neglect of horizontal diffusion. Inclusion of these effects will round off the ridge crest. In general the topography about oceanic rifts lies between the predictions of the narrow dyke intrusion model of rifting, included in the plate model, and those of the simple extensional rift model presented here. Nevertheless both the heat flow and topography at ocean ridges suggest the presence of a broad zone of intrusion below the ridge crest.

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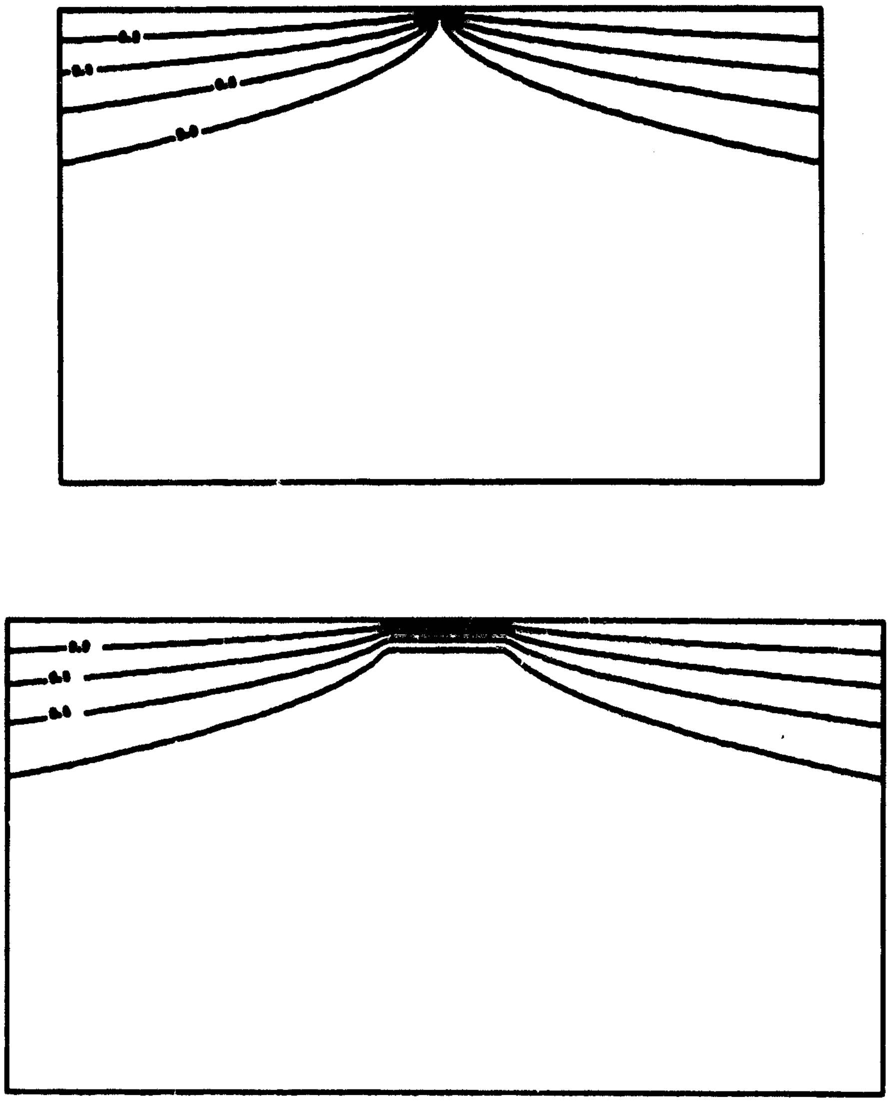

Comparison of Ridge Model Isotherms. The upper figure shows isotherms in the vicinity of a mid-ocean ridge according to McKenzie's plate model. The lower figure shows isotherms for an extensional model with $G^1 = 410$. Isotherms are shown for $T = 0.2T_1$, $0.4\ T_1$ $0.6T_1$ and $0.8T_1$ in both figures. The upper and lower surfaces in each case also represent the $T = 0$ and $T = T_1$ isotherms respectively.

LITHOSPHERIC THINNING OVER MANTLE HOTSPOTS: A MECHANISM FOR RIFTING CONTINENTS. S. Thomas Crough, Department of Geosciences, Purdue University, West Lafayette, IN 47907

As pointed out by several investigators, there is a significant correlation between continental rifting and the formation of broad domal uplifts. Modern grabens such as the Ethiopian and East African rifts cut through major plateau areas; older rifts are often associated with major coeval unconformities, suggesting that they were also formed contemporaneously with regional uplift. It has been suggested that rifting usually proceeds in the following sequence of events: 1) swells form in a continental region, 2) swells crack and develop radiating grabens, and 3) grabens on neighboring swells link up to form continent-wide rifts. Recent improvements in our understanding of how swells form lend support to this proposed scenario by demonstrating its physical plausibility.

A variety of morphological, geological, and geophysical data imply that the midplate swells associated with hotspots are uplifted by a broad-scale reheating and thinning of the lithosphere. This thinning mechanism satisfactorily explains: 1) the persistance of swells beneath old hotspot traces, 2) the gradual subsidence of swells in time, 3) the positive gravity and geoid height anomalies measured over swells, and 4) the heat flow high along the crest of the Hawaiian Swell. While other possible support mechanisms are consistent with one or two of these observations, no other mechanism is consistent with all of them.

Lithospheric thinning can cause rifting in two ways: firstly, it creates a zone of weakness by cutting plate thickness to about one-third its original amount, and secondly the consequent uplift generates tensile stresses sufficient to fault pre-fractured rock. The importance of the first mechanism, reducing plate thickness, depends on the long-term creep properties of the lower lithosphere. If the bottom of the plate is sufficiently ductile then it has no strength and removing it will not alter plate strength. Theoretical calculations and inferences from flexural rigidity studies suggest that lithospheric thinning does not substantially affect plate behavior. The uplift effect is the dominant rifting mechanism. Calculations similar to those used to estimate the ridge-push at spreading centers show that swell-push can generate deviatoric tensile stresses on the order of several hundred bars.

The major problem with the swell-rift model is not how to generate rifts from uplift, but how to explain the lack of rifts on most swells. Given the present population of approximately 40 hotspots and associated swells, it is perhaps puzzling how the plates manage to remain largely intact. Three possible resolutions of this paradox can be suggested. First, the maximum tensile stresses caused by swell uplift never exceed the magnitude of the compressive stresses generated within oceanic lithosphere by ridge push. Therefore the crests of oceanic swells should remain in deviatoric compression and oceanic swells should rarely rift. Only swells formed on continental lithosphere generate deviatoric tensile stresses. Second, the magnitudes of these continental stresses are dependent on swell height. As noted by earlier workers, continental swells are more fully developed in younger orogenic areas than in older cratons. Therefore, continental rifting may require a fortuitous alignment of hotspots beneath younger terrains. Third, the deviatoric stresses within the lithosphere depend on plate geometry and motion as well as on long-wavelength topography. Because these other stresses are also estimated to be in the hundred bar range, they may, in places, negate the tendency of swells to rift.

THE OSCILLATORY (NON-STEADY-STATE) MODEL OF THE ORIGIN AND DEVELOPMENT OF THE OCEANIC AND CONTINENTAL RIFTS

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From the time of the discovery of mid-oceanic ridges, alternative models have been proposed for their origin. At first ridges were considered to be the topographic expression of the ascending limb of a convection cell in the mantle (Turcotte and Oxburgh, 1967), then as a result of the cooling of rigid lithosphere plates moving away from the place of their formation (Langseth et al., 1966; McKenzie, 1967). M. Talwani et al. (1965) and A. Grachev (Grachev and Karasik, 1966; Grachev, 1968, 1972, 1977) have advanced a model to account for mid-oceanic ridge generation with anomalous mantle. In this model the hot and low density anomalous mantle gives rise to the uplift of the oceanic crust (as with continental rifts).

In the past decade, the steady-state Langseth-McKenzie model has been widely accepted. This model predicts that the depth of the mid-oceanic ridges decreases as $t^{1/2}$, where t is the age of oceanic lithosphere and the depth of the ridge crests is a constant value (about 2700 m) in all cases and independent of the spreading velocity (Sclater et al., 1971; Parsons and Sclater, 1977, and others). E. Schneider and P. Vogt (1968) were the first to show that there are significant differences between the steady-state model predictions and actual ridges. Indeed, the depth of the ridge crests varies significantly from one place to another one within the same mid-oceanic ridge, and from ridge to ridge in the World Rift system (Grachev, 1976, 1977). In these cases for every deviation from the model it is necessary to search for a corresponding explanation (for example, the introducing of a third type of spreading ridge, the hot spot ridge, etc.). However, the main obstacle to the simple cooling model is an absence of a satisfactory mechanism to explain systematic deviations from the predicted depths in "critical regions" of the oceanic crust, where the age of the floor is greater than 80 m.y. B.P. (Parsons and Sclater, 1977). There are also a great deal of other data from ocean geology which are beyond the framework of a plate tectonic model. Among the supplementary data, the most important phenomenon is a recurrent interchange between eustatic transgressions and regressions during the last 200 m.y. which may only be explained by a depth change of all mid-oceanic ridge systems. All available data suggest that the mid-oceanic ridges are not steady-state features of the ocean floor topography.

It is very important to underline that continental rifts have a similar dependence of the rift shoulder heights versus distance from the rift axis to oceanic rifts. Moreover, both continental and oceanic rifts have a limit of uplift relative to the adjacent undisturbed crust (Grachev, 1972, 1977). We can conclude that there is a common reason responsible for these relationships in both cases.

As a base for our model of the rifting mechanism we have taken the idea of a recurrent input to the high level of the lithosphere of hot anomalous mantle material. The latter is a result of density differentiation of the lower mantle at the core-mantle interface (Artyushkov, 1968, 1970, 1971, 1979). The time span of such cycles is about 25-35 m.y. Computer modelling has shown that independent of the anomalous mantle viscosity, there are three stages in rift development. These stages reflect the lateral spreading of the anomalous mantle at the base of both continental and oceanic lithosphere.

During the first stage, the injection of a low density and hot anomalous mantle layer beneath the crust produces ar isostatic uplift. The second stage corresponds to the fast lateral spreading of the anomalous mantle, and the extension of the uplift area. It is important that the rate of the mantle spreading strongly decreases for the time span of about 10^6 years. As a result, the anomalous mantle lens freezes and isostatic uplift ceases. At the beginning of the third stage the slow subsidence commences due to the cooling of the anomalous mantle.

The character of the curves depicting the spreading of the anomalous mantle under the mid-oceanic ridges and continental rifts depends on mantle viscosity, but the thickness of the anomalous mantle lens can produce the observed variations in the depths of ridge crests (fixed Sclater constant) and the heights of the continental rift shoulders. Both values (viscosity and uplift) also depend on degree of the partial melting of the anomalous mantle which can vary from 1 to 10-15%. Together the data allow us to explain the actual variations of the mid-oceanic ridges depths and do not require us to solve the problem of the "critical depths" for the age greater than 80 m.y. B.P.

It is easy to show that the basic shape of the cumulative curve for all three stages of rifting should be characterized by two limits of saturation pertaining to the very beginning of the first stage and to the ending of the third stage. Thus, the proposed scheme has the same trend for each cycle of mid-oceanic ridge development as a steady-state model but the starting point (initial depth of the ridge crest) varies with the thickness change of the anomalous mantle lens.

The depicted process of both oceanic and continental rifting could operate more than once in the Earth's history in time of input of the anomalous mantle. The global epochs of folding, volcanism and uplifts of the continents and synchronized periods of activity of the mid-oceanic rift system can be explained within the framework of the oscillatory rifting model.

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THERMOMECHANICAL MODELS OF DEVELOPMENT OF CRATONIC SEDIMENTARY BASINS BY RIFTING

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Some sedimentary basins on old cratonic crust appear to be located over the centers of ancient rifts. For example, the nearly-circular Michigan and Williston Basins, the Illinois Basin, and the Rome Trough, West Virginia, all reveal distinct positive Bouguer anomalies similar to those over present rift areas. The subsidence history and geometry of these basins also suggest that a load caused by the diapiric intrusion of subcrustal material may have caused the subsidence. In young rifts, uplift, diapiric intrusion, and graben formation is commonly followed by surface subsidence over the central axis. The development of these basins over rifts may be, in part, caused by the viscoelastic response of the lithosphere to a diapiric mass excess within the crust. Study of this problem requires a thermomechanical model that is sensitive to temperature-variation and assumes a non-linear rheology.

In these models, we consider continental rifts that have the following features: (i) A negative Bouguer anomaly about 450 km wide that contains a strong positive Bouguer anomaly about 50-75 km wide over the axis of the central graben (Fairhead and Girdler, 1972). The positive anomalies are associated with mafic igneous intrusions. (ii) Seismic studies and the presence of pervasive normal faulting indicate that near-surface stress is mainly tensional in rift environments, with the principal axis of tension oriented perpendicular to the rift axis (Fairhead and Henderson, 1977). (iii) A zone of high geothermal gradients, symmetrically disposed about the central graben, indicates the presence of high heat-flow localized over a central graben (Morgan and Wheildon, 1981).

We have formulated a model of lithospheric flexure according to a viscoelastic extension of Euler-Bernoulli beam-column theory. The lithosphere is represented rheologically as a viscoelastic material, incorporating both a linear elastic strain component and a non-linear viscous dislocation-glide creep component. We employ the non-linear Maxwell constitutive relation

$$\frac{\partial \varepsilon}{\partial t} = \frac{1}{E}\frac{\partial \varepsilon}{\partial t} + A\,\sigma^n + \alpha \frac{\partial}{\partial t}(\Delta T),$$

where ε is the axial strain, σ is axial stress, E is Young's modulus, α is the coefficient of linear expansion, ΔT is the transient temperature variation, A is the inverse viscosity coefficient, and t is time. The first term on the righthand side governs the elastic strain component; the second governs the dislocation-glide creep component; and the third term governs the effects of thermal expansion. The thermal expansion causes dilitation (volume change) which in turn produces density variation and gravitational instability (buoyancy). The inverse

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viscosity coefficient is very sensitive to temperature and is of the form

$$A = \dot{\varepsilon}_o \frac{\Omega}{KT\mu^{n-1}} \exp\left[\frac{-\left(gT_m + \frac{PV^*}{K}\right)}{T}\right]$$ (Ashby & Verall, 1977; Vetter & Meissner, 1977)

where T is absolute temperature, P is lithostatic pressure, V^* is the activation volume for dislocation-glide creep, T_m is the solidus temperature of the upper mantle, K is Boltzman's constant, and g, $\dot{\varepsilon}_o$, Ω, and μ are material constants. P varies with depth; T is taken as a function of depth, of lateral position, and of time if transient temperature effects are considered.

Vertical forces (loads), which are similar to geologically occurring gravitational instabilities within the lithosphere, are applied to the beam. The character of these loads over rifts is suggested by the two prominent gravimetric anomalies. The broad Bouguer low is attributed to a mass deficiency deep in the lithosphere and produces a bouyant upward force; a positive load is associated with the sharp Bouguer high and is attributed to a mass excess (basic intrusion) high in the lithosphere. The net load is superposition of these two loads. For the cases considered in this model, this net load consists of a central positive zone, 50-150 km wide, flanked on either side by negative zones. The anomalous heat flow associated with the central rift suggests high temperature gradients which significantly increases creep and deflection rates in this region.

Solutions to this model reveal deformational characteristics that seem to link the mafic intrusive to basin formation. The deflection history consists of an initial elastic response to the crustal intrusive load followed by a long period of downward creep. Initially balancing this effect, the buoyancy caused by the broad mass deficiency in the lower lithosphere produces a symmetric upwarp about 400-600 km wide; this has a central maximum amplitude of 0.7-1.5 km. Higher than normal local stress surrounding the central intrusive causes accelerated viscous effects that create a progressive flattening of the central elevated dome region, so that after approximately 10^6 yr, a central depression 50-150 km wide develops. If the mass deficiency associated with the buoyant load is attributed to a transient thermal event deep in the lithosphere, the upwarp is reduced as cooling take place; this eventually leaves an isolated "basin" in the central region after the thermal disturbance dissipates.

In these evolutionary models of a rift, stresses at depths greater than 50-60 km decay rapidly, providing a lower boundary for the effective mechanical lithosphere. The stress-decay character of the lithosphere is locally accentuated in the zone of elevated geothermal gradients, where an effective thickness of the lithosphere is less than 35 km under the central depression. The broad upwarp is characterized by tensional stresses above 30 km in the lithosphere. Interaction of stress due to the sinking central high density mass and stress from the broad upwarp yield in some examples shear stresses that may be sufficient to produce the rift grabens.

The Bouguer anomalies, corrected for sedimentary rock attraction, over the Williston (Datondji, 1981) and Michigan basins(Haxby, Turcotte and Bird, 1976) show positive anomalies up to 90 mgals centered over the maximum thickness of the basins. The anomalies are attributed to diapiric intrusions in the crust, perhaps during a late P€ rifting event. Subsidence history of these basins suggest, however, that a period of regional stability preceeded the basin formation in the early Paleozoic and no uplift is evident in the stratigraphic record just prior to formation of the basin. These data suggest that a diapiric intrusive mass may have localized the center of subsidence. The process that seems to have accelerated the viscoelastic subsidence of the mafic mass in the lithosphere during the early Paleozoic subsidence phase is not well understood.

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CONVECTIVE THINNING OF THE LITHOSPHERE: A MECHANISM FOR THE INITIATION OF CONTINENTAL RIFTING. T. Spohn and G. Schubert (both at: Dept. of Earth & Space Sciences, Univ. of California, Los Angeles, CA 90024).

Active continental rifting is an important element of the well-known Wilson cycle of tectonics; it precedes continental breakup and the opening of a newly formed ocean. Modern examples of continental rifts include the East African Rift Zone, the Baikal Rift, the Rhinegraben Rift and the Rio Grande Rift. All continental rift zones are characterized by anomalously high heat flow, long wavelength Bouguer anomalies and anomalous seismic structures. Together these observations point to anomalously thin lithospheres beneath continental rift structures.

Thinning of the lithosphere by heat carried convectively to its base is a possible mechanism for the initiation of continental rifting. The lithosphere-asthenosphere boundary marks the transition from a quasi-rigid behavior of the mantle on geological time scales to a viscous one. The boundary lies at or close to the solidus of the mantle material and its position is a function of temperature and pressure. Additional heat carried to the boundary will cause it to rise and the lithosphere to thin. The increase in heat flux from the asthenosphere might be caused by the emplacement of a plume, for example. Using a one-dimensional model, we show that these plumes can rapidly thin very thick lithospheres while not perturbing the surface heat flow extensively.

We consider a horizontally infinite continental lithosphere across which temperature T varies with depth z initially according to

$$\begin{aligned} T(z) &= \frac{q_i}{k} z + A(z) = \frac{q_i}{k} z + \frac{A_o D^2}{k}\left\{1 - \exp(-\frac{z}{D})\right\} , && 0 \le z \le z_o \\ &= \frac{q_i}{k}(z - z_o) + T_o(z_o) , && z_o \le z \le \ell_o \end{aligned} \tag{1}$$

where z_o is the thickness of the crust, ℓ_o is the thickness of the lithosphere, q_i is the heat flux from the mantle, and D is a characteristic length for the depth distribution of heat sources. The quantities q_i, A_o, and D are related to the surface heat flux q_s by

$$q_s = q_i + DA_o \quad . \tag{2}$$

At time t = 0 we suddenly increase the heat flux into the lithosphere by placing a plume beneath it, for example. A stagnation point boundary layer develops and the heat flux into the lithosphere is increased to q. Thus, thermal equilibrium is perturbed and the lithosphere-asthenosphere boundary starts to rise to a new equilibrium position. The space left by the rising boundary will be immediately filled by the plume, i.e. the thermal boundary layer follows the lithosphere upward. Both the transition temperature $T_\ell(z)$ and the plume temperature $T_p(z)$ can vary linearly with depth. Within the lithosphere the heat conduction equation is valid and it is subject to the boundary conditions

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$$T = 0 \text{ at } z = 0 \quad , \quad T = T_\ell(z_\ell) \text{ at } z = z_\ell(t) \quad , \qquad (3)$$

where $z_\ell(t)$ denotes the base of the lithosphere as a function of time.

Energy conservation at the lithosphere-asthenosphere boundary provides an equation for its motion. The heat consumed on transforming lithosphere to asthenosphere is balanced against the heat fluxes to and away from the boundary

$$\left\{ \rho L + \rho c (T_p(z_\ell) - T_\ell(z_\ell)) \right\} \frac{dz_\ell}{dt} = -q + \left(k \frac{\partial T}{\partial z}\right)_{z=z_\ell} \quad , \quad z_\ell \geq z_o \ . \quad (4)$$

In (4) L is the latent heat (if applicable) and k is the thermal conductivity. Equation (4) together with the heat conduction equation and the initial and boundary conditions constitutes a nonlinear boundary value problem similar to the classical Stefan Problem. We have solved this problem numerically using a finite difference method.

The ratio of the amount of thinning to the initial thickness of the lithosphere is directly proportional to the increase in heat flux from the asthenosphere and inversely proportional to the difference between the final subcrustal geothermal gradient and the slope of the transition temperature profile. The rate of thinning is determined by the initial and final equilibrium thicknesses, the thermal diffusivity, the transition temperature profile and the plume temperature profile. For incipient melting, the rate of thinning depends only slightly on any latent heat associated with the lithosphere-asthenosphere transition. During a large fraction of the time between the initial and final equilibrium states the lithosphere thins at a rate which is inversely proportional to the square root of time t.

Thick lithospheres ($\ell_o \approx 300$ km) can be thinned to the crust for a 5 times basal heat flux enhancement γ. However, γ must be larger than 10 if this is to occur on time scales of tens of million years or within 100 million years. Thinner lithospheres ($\ell_o \leq 150$ km) can be thinned on that time scale for $\gamma \leq 5$. Large increases in heat flux from the asthenosphere ($\gamma > 5$) which are needed to thin thick lithospheres on reasonable time scales or even to thin the 150 km thick lithosphere to the crust in a few tens of million years are compatible with estimates of the heat flux advected by mantle plumes. The excess heat carried to the base of the lithosphere by the plume eventually reaches the surface via conduction in this model. Because radioactive heat production in the crust is constant with time, the increase in surface heat flow is much less than the increase in basal heat flow. Many of our models that thin the lithosphere in reasonable times have final equilibrium surface heat flows less than or equal to the upper maximum continental heat flow of about 150 mW m^{-2}. However, the observed surface heat flow in a rift zone may never, in fact, fully reflect the enhanced basal heat flow because surface heat flow lags behind lithospheric thinning and the time lag increases with γ. Furthermore, differentiation of basaltic magma from the plume at crustal level or below 50 km say, might consume the energy advected by the plume and the heat might eventually reach the surface via magmatism and volcanism.

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We have also calculated uplift due to thermal expansion during thinning and get values of the order of 10^3 m. These uplifts should be sufficient to explain observed uplifts at sites of continental rifts. For example, the uplift for the lower Rhine Rift is estimated to be about 600 m. While our model of convective lithosphere thinning does not account for surface volcanism and continental breakup, it is capable of explaining such geological observables as anomalous lithosphere thicknesses, surface heat fluxes, thinning rates and uplift rates.

THERMAL EXPANSION AND UPLIFT OF THE LITHOSPHERE PRECEDING RIFTING.

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Broad uplifts of the continental crust accompany rifting episodes. One hypothesis is that these uplifts result from the stretching of the lithosphere which is the cause of passive asthenospheric diapirism and heating of the lithosphere. An alternate hypothesis is that heating of the lithosphere, lithospheric thinning and uplift are the cause of rifting. The latter hypothesis has been examined in detail and several mechanisms for the heating of the lithosphere and resulting uplift have been investigated: (1) the conductive heating caused by an increase of the heat flow or the temperature at the base of the lithosphere; (2) the convective heating caused by the injection of magma into the lithosphere; and (3) the active diapiric uprise of the asthenosphere into the lithosphere.

The thermal expansion of the lithosphere caused by conductive heating has been studied analytically. It does not appear to be an adequate model of uplift and lithospheric thinning (at least as far as rifting is concerned) for two reasons: the process is too slow and requires an extremely large thermal anomaly. Although the uplift starts instantly after the initiation of the thermal anomaly at the base of the lithosphere it will be completed in a time of the order of 100 10^6 years. Furthermore, a fairly large thermal anomaly at the base of the lithosphere is required to produce the amplitude of uplift observed.

The heating of the lithosphere following convective heating by the injection of magma has been modeled by assuming that vertically moving heat sources are introduced into the lithosphere. Initially, the uplift velocity increases linearly with time and is proportional to the intensity of the heat sources and to their vertical velocity. This mechanism is a more satisfactory explanation for the uplift because most of the uplift can be completed in a geologically short time (less than 5 10^6 years) provided that the vertical heat source's velocity is large enough. Some difficulty remains with this mechanism, however, because a very high heat source intensity is needed to produce the amplitude of uplift observed in rift zones: this implies either exceedingly high temperatures for the rising magma or a large quantity of magma.

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Mareschal, J.-C.

The most satisfactory mechanism for the initiation of rifting appears to be the diapiric uprise of the asthenosphere into the lithosphere. This can take place only if: (a) a gravitational instability is present between the lower lithosphere and the asthenosphere; (b) the effective viscosity of the lower lithosphere is sufficiently low ($\sim 10^{21}$ decapoises) for the process to be completed in a geologically short time. It must be assumed that heating of the lower lithosphere has reduced its viscosity before the asthenopheric diapir can rise. The heating mechanism could be conduction or convection by magma injection. Preliminary numerical models of the conductive heating of the lithosphere and of the following diapiric uprise of the asthenosphere, assuming a temperature dependent newtonian viscosity, have been investigated. The models show that, once the heating of the lower lithosphere has reduced its viscosity to about 10^{21} decapoises, the diapiric uprise of the asthenosphere could cause the lithospheric thinning and most of the accompanying uplift to occur in a time of the order of 10 10^6 years.

GANYMEDE TECTONICS: GLOBAL SCALE RIFTING DUE TO PLANETARY EXPANSION?,
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Voyager images of Ganymede show clear evidence for global tectonic activity (1,2). The surface consists of about equal proportions of heavily cratered, dark terrain and younger, less cratered bright terrain. Dark terrain, thought to be an ice-silicate mixture, generally forms polygonal regions separated by bright terrain as shown in Figure 1. Bright terrain occurs in long, curvilinear bands that divide dark polygons, wedge-shaped regions that partially divide polygons, and irregularly shaped regions that may form by the intersection and coalescence of bands. It is composed of relatively clean ice. The emplacement of bright terrain by the extrusion of a relatively water-rich magma in regions of lithospheric extension is geologically reasonable and is supported by a variety of photogeologic evidence.

Three different styles of extensional tectonics which may be associated with the emplacement of bright terrain are a.) tension fracturing and surface flooding with minimal separation between blocks of dark terrain; b.) lithospheric spreading involving lateral motion of blocks of dark terrain; and c.) rifting due to finite lithospheric extension. These simple models provide a conceptual framework for interpreting a variety of geologic data. Dark/bright terrain contacts are usually linear and sharply defined. Craters and other features transected by a contact are sharply truncated. Although flooding of dark terrain has occurred in some areas, particularly around irregular regions of bright terrain, bright material, especially in bands of bright terrain, appears to have been structurally confined.

Several types of observations distinguish between lithospheric spreading (b) and finite lithospheric extension (c). If the former has occurred, the two parts of a crater transected by bright terrain should appear on opposite sides of the band. This has not been observed. Since bright terrain covers a large fraction of the surface, its creation by lithospheric spreading requires large-scale destruction of dark terrain. If all of the bright terrain is formed by lithospheric spreading, compressional deformation should occur in the dark terrain. As discussed below, none has yet been observed. Photogeologic evidence therefore favors the emplacement of bright terrain by rifting associated with finite lithospheric extension.

Several structural styles of rifting have been suggested. Dark terrain may have subsided along normal faults, as in terrestrial rift zones, forming graben (3) or may have stoped (4) or foundered (5) into the deeper interior. The occurrence of small, isolated regions of dark terrain in irregularly shaped areas of bright terrain argues against stoping. Piecemeal stoping would not be consistent with the continuity and linearity of bright band edges. Although these observations favor the formation of graben, the thickness of bright deposits derived from detailed geologic mapping of the albedo of material excavated by impact craters and the burial of pre-existing topography will provide more information on the structure of rift zones.

Further evidence for the style of global tectonic evolution can be obtained by an understanding of individual structural features having the topographic form of narrow, linear depressions, collectively called grooves (1,2). The simplest examples are found in the dark terrain where grooves occur singly and in nearly parallel pairs with up to about 10 km spacing between individual grooves. Groove pairs, as shown in Figure 2, commonly merge along strike to form single grooves with widths near the resolution limit of Voyager images. Groove pairs are interpreted to be viscously relaxing graben (3). If graben viscously relax in a manner similar to that recognized for impact craters (6,7), doming of the graben floor will produce a pair of depressions similar to the groove pairs observed on Ganymede. At their present resolution, single grooves may be either extension fractures or narrow graben. No compressional features having the topographic form of thrust fault scarps or fold trains have yet been recognized in the dark terrain.

Grooves are more abundant in the bright terrain where they occur as single grooves, groove pairs, and in sets of multiple parallel grooves. Single grooves and groove pairs appear similar to those in the dark terrain. Many groove sets can be constructed by placing single grooves or groove pairs side by side. Although this argues for an extensional origin, the nearly sinusoidal topography of many groove sets could also be caused by the folding of a more viscous, near-surface layer. Thus a compressional origin for many groove sets can not yet be ruled out.

Since Ganymede may have undergone relatively large changes in volume (8), progressive fragmentation of the lithosphere by membrane stresses during planetary expansion would be an attractively simple mechanism for explaining global rifting. To test this hypothesis, the lithosphere of Ganymede may be treated as a thin viscoelastic shell. A near-surface viscosity of 10^{25} poise would allow 10 km wide topographic features to relax in times on the order of 10^9 years. Thus, the relaxation time η/G of a Maxwelliam material with a shear modulus $G=10^5$ bars is only 3×10^6 years so that viscous relaxation of elastic stresses is likely to be important. The stresses within an unbroken dark region can be calculated using membrane tectonics (9) modified for a viscoelastic shell. The largest deviatoric stress in a circular shell of angular radius ϕ_0 is the azimuthal or hoop stress σ_θ at the edge of the shell. For a shallow shell and a rate of expansion that is relatively uniform on the relaxation time scale,

$$\sigma_\theta = 2\eta\phi_0^2\dot{R}/4R + \sigma_{\phi 0}$$

where R and $\dot{R}$ are the planetary radius and its time rate of change, and $\sigma_{\phi 0}$ is the stress applied to the edge of the shell. This simple result shows that planetary expansion will create tensional hoop stress, and therefore extensional failure, at the edge of an unbroken dark shell. This may explain the wedge-shaped regions of bright terrain that extend into dark terrain.

The magnitude of the hoop stress also increases with the size of the shell. Therefore, in a simple model with an initially uniform lithosphere and vanishing boundary stress, unbroken regions of comparable size would be expected at any stage of fragmentation. This appears to be contradicted by the wide range of sizes observed. If global scale rifting is to be explained by planetary expansion alone, an initially nonuniform lithosphere and/or the transmission of stresses across regions of bright terrain must be important. Given the complexity and diversity of grooves in regions of bright terrain, significant differences in strength of such regions would not be surprising.

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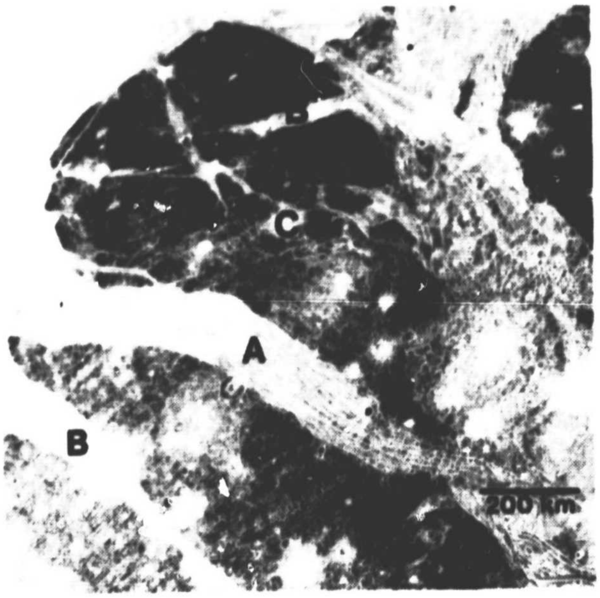

Figure 1. Polygonal regions of dark terrain separated by bands of bright terrain containing groove sets. Wedge-shaped regions of bright terrain are also shown.

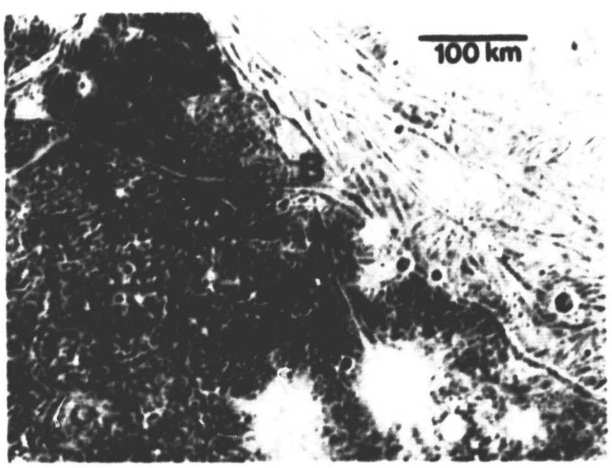

Figure 2. Example of groove pairs in the dark terrain. Each groove pair narrows along strike to form a single groove. Grooves in a complex area of bright terrain are also shown.

VENUS: GLOBAL-SCALE CRUSTAL DISRUPTION, INCLUDING RIFTING OF CONTINENTAL ROCKS - Gerald G. Schaber, U.S. Geological Survey, Flagstaff, AZ 86001.

It is clear from analysis of Pioneer-Venus spacecraft radar altimetry data that Venus does not currently possess the same style or magnitude of plate tectonic activity that exists on Earth, although both bodies have essentially the same diameter and density. Venus does not possess deep, "oceanic" lowlands with central rises and rifts, or "continental" masses with well-defined marginal troughs (Masursky et al., 1980; Phillips et al., 1981). The question remains, however, whether Venus ever did undergo global-scale, crustal or lithospheric disruption of any kind, and if so, whether we can recognize evidence for such events on its surface, given the resolutions presently available from Pioneer-Venus and Earth-based radar data. The answers to both questions may be yes; Venus' surficial topography, surface slope distribution and gravity perturbations may indicate extensional rifting and associated volcanism along ancient, global scale fractures within a substantially thick, continental crust (Fig. 1).

Several authors (Malin and Saunders, 1977; Masursky et al., 1980; McGill et al., 1981) have described evidence, based on both Pioneer-Venus and Earth-based radar data, of dominantly continental-style rifting that trends north-south between Beta and Phoebe Regios. The analogy given between the long Beta-Phoebe rift zone (approximately 6,000 km long) and its associated elevated terrains, and the afro-Arabian rift zone (6,500 km long) and associated Kenyan and Ethiopian uplifts is made believable by the similar scale.

At this conference I will expand earlier discussions (Masursky et al., 1980; Schaber and Masursky, 1981a,b) and concentrate on two additional regions where "continental" crustal disturbances may have occurred on Venus. These have no terrestrial counterpart in scale, but are of the same magnitude as the Earth's pervasive mid-oceanic rise and rift systems. A general pictorial overview of extensional rifting on smaller bodies in the solar system will also be given, if time permits.

The region of Venus bounded by lat 40° N. to 43° S. and long 60° to 300° contains complex ridge-and-trough terrain and associated upland regions that have anomalously high rms slopes and centimeter-scale surface roughness values. Included are three broad linear zones that are centrally located along the crests of extremely broad, low rises. Area of the disrupted zones is about 4.6×10^7 km^2; that of their associated rises is approximately 20% of Venus' surface area, or 9.2×10^7 km^2.

The longest tectonically disrupted zone, which I will call the Aphrodite-Beta linear, trends northeast between the south-facing slopes of Aphrodite Terra to the west side of Beta Regio. It passes through Atla Regio, an elevated region centered at lat 4°, long 200° (Fig. 1). This extensive surface linear is actually composed of discontinuous regions where tectonism was more intensive; taken as a whole it extends halfway around the planet (20,000 km) on a broad, low rise.

The regions south and east of Aphrodite Terra are dominated by well-developed ridge-and-trough topography that lies upon a low (1.5 km above mean planetary level) plateau whose area is about 3×10^7 km^2 (Figs. 1,2). The ridge-and-trough features are both linear (Dali Chasma) and curvilinear (Artemis Chasma) in plan; the troughs average about 160 km in width and sometimes exceeds 2.5 km in depth. The trough widths may be overestimated and trough depths underestimated because the P-V altimeter footprint is large. Ridges marginal to the troughs are commonly unequal in height where they occur on opposite sides of the depressions; they rise from 0.5 to 2.5 km above the surrounding plateau. The origin of the marginal ridges may be related to both volcanism and isostatic adjustment. The deepest and best developed of the ridged troughs, within the Aphrodite-Beta linear, are intimately associated with the south-facing slopes of Aphrodite Terra and the plateau to the east through Diana and Dali Chasmata, ending at lat 15° N., long 182°. Increased rms surface slopes for the ridge-and-trough features suggest that the walls may be extremely rough.

The northeast end of the Aphrodite-Beta linear passes through Asteria Regio; it consists of discontinuous regions where rms slope and meter-scale surface roughness values are increased. It extends southwest from Beta Regio a distance of about 8,500 km, intersecting Atla Regio (Fig. 1). The zone of disruption exceeds 2000 km in width.

The second major zone of surface disruption within this region, which I will call the Themis-Atla linear, extends 14,000 km, just northwestward from east of Themis Regio (lat 40°, long 297°) to the northeastern tip of Aphrodite Terra at lat 25° N., long 185°. It intersects the N-E trending Aphrodite-Beta linear as it passes through Atla Regio. Atla Regio is positioned at the intersection of both global scale linears; this is considered strong evidence for a volcano-tectonic origin. The Themis-Atla linear is not well defined by Pioneer-Venus rms-slope data; it is recognizable in the altimetry data. Northwest of Atla Regio, the Themis-Atla linear appears to be offset to the east, where it is composed of several linear troughs on the crest of a linear ridge that is as much as 3.5 km higher than the mean planetary level. The ridge, which forms the northwest end of the Themis-Atla linear, is thought to represent major eruptive volcanism sourced in this northwest-trending crustal disruption.

The Beta-Phoebe, Aphrodite-Beta and Themis-Atla linears shown on Fig. 1 are thought to reflect ancient disruptions of sialic continental crust; the ridge-and-trough complexes associated with the Beta-Phoebe linear and Aphrodite Terra, may represent relatively recent uplift and extension.

The absence of true "oceanic" lowlands that contain medial rises and rift valleys, the assumed anhydrous condition of the lithosphere, inferences that the crust is thick, and high surface temperatures (450°C higher than Earth's), are indications to some workers that early "continental-type" rocks may be globally distributed, and a subsequent "choking-off" of continental disruption, such as plate tectonics, resulted from early, rapid thickening of a volatile poor lithosphere (Schaber and Boyce, 1977; Kaula, 1980; Anderson, 1981; Phillips et al., 1981). This appears to be the case.

Arvidson and Davies (1981) and Head et al. (1981) have properly questioned whether we can directly compare topographic maps of the Earth and Venus in a search for comparable morphologic evidence of plate tectonics. Such studies have shown that it is difficult to verify the known existence of plate tectonics on Earth when resolution is reduced to that of the Pioneer-Venus altimetry, much less verify it on a planet like Venus, where such processes are not known to exist. Recognition of plate tectonic-like features on Venus may be hindered or at best complicated by the general broadening and attenuating of elevated features associated with spreading centers that may exist, resulting from high contrasts in density and small differences in temperature between the surface and the fully convecting interior (Arvidson and Davies, 1981; Head et al., 1981). Adopting these assumptions, Kaula and Phillips (1981) established quantitative tests for

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plate tectonics on Venus. Of the four tests they described, two appeared to be satisfied in the Pioneer-Venus altimetry data: (1) the existence of a broad reference plains where variations in elevation are small, and (2) ridges that are predominantly concave in shape rise above the plain. Kaula and Phillips (1981) concluded that the upper limit on the rate of plate creation on Venus is about 25 percent that of the Earth, and this in turn limits to 15 percent the heat delivered to the surface of Venus by convection; on Earth 70 percent is convected.

The presence on Venus of extensive, subdued rises with centrally located surface disruptions is thus considered to be strong evidence for extensional tectonics and associated volcanic extrusion. These are influenced by thermal pulses in the asthenosphere. Such pulses convect mass and heat upward, to produce regional doming, rifting with associated marginal ridges, and localized crustal thinning. Based on terrestrial ratios of rift widths to crustal thickness (Meinesz, 1950), the venusian rifts, which range from 125 to 200 km in width, would predict a crustal thickness of 75 to 150 km below the rises (Schaber and Masursky, 1981a).

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Fig. 1- a) Shaded relief map of Venus derived from Pioneer-Venus radar altimetry; b) sketch geologic map of global surface disruptions on a region of Venus derived from analysis of Pioneer-Venus altimetry, rms slope data and radar image data;c) topographic map of a region of Venus showing the location of broad, low rises associated with the disruption linears shown in b).

Fig. 2- 3-dimensional isometric diagrams showing the cross sections of ridge-and-trough features at a) Artemis Chasma and b) Diana and Dali Chasmata

RIFTING IN WESTERN AND CENTRAL EUROPE

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Rifts play a pre-eminent role amongst the sedimentary basins of Western and Central Europe. During Late Palaeozoic to Cenozoic time several more or less distinct rifting cycles are recognised whereby graben formation took place in a number of different magatectonic settings (Ziegler 1981).

Rifting leading to the break-up of continents and the opening of major oceanic basins, for instance on the scale of the Atlantic, is the most important process of graben formation. The driving mechanism of such mega-rift systems may well be convection currents in the asthenosphere which excert tensional stresses on the lithosphere and ultimately cause its failure (Richter and McKenzie 1978, Chase 1979, McKenzie et al. 1980). Regional crustal extension preceeding continental splitting can affect wide areas around future plate boundaries. With the progression of crustal extension a polarisation of the rift systems can often be observed whereby peripheral grabens become inactive whilst principal rifts remain active until crustal separation takes place (e.g. Mesozoic rift system of Western and Central Europe). In such a tectonic setting the duration of the rifting stage preceeding crustal separation can be highly variable. For instance in the Central Atlantic it lasted only some 40 Ma. while in the Norwegian-Greenland Sea it lasted about 270 Ma.

Back-arc rifting is thought to have played a significant role during the Devonian and Early Carboniferous development of the Variscan geosynclinal system (Ziegler 1982). The principle mechanism governing back-arc rifting and sea-floor spreading is thought to be secondary convection currents in the mantle wedge immediately above the subducting lithospheric slab. Tensional stresses excerted by this convective system on the crust of the overriding plate are apparently only then able to induce back-arc rifting and sea-floor spreading when the convergence rate between the subducting and the overriding plates is relatively low or if they diverge. This may be accompanied by partial decoupling of the two plates at the Benioff zone which, under these conditions is likely to display a relatively steep dip (Marianas type setting). On the other hand, if convergence rates are relatively high, stronger coupling at the Benioff zone results in compressive deformation of the back-arc areas and the overpowering of the back-arc convenctive system. Under such conditions the Benioff zone is likely to dip relatively gently under the overriding plate (Andean type setting), (Uyeda 1981, Hsui and Toksöz 1981, Zonenshain and Savostin 1981). As convergence rates between plates are not constant but appear to change through time periods of back-arc extension and compression can alternate with each other. Correspondingly back-arc rifts and oceanic basins are prone to destruction, but particularly under the impact of continent-arc collision. This is exemplified by the Late Visean to Early Stephanian Variscan orogeny during which the Central Armorican Saxothuringian and the Cornwall-Rhenish-East Sudetic back-arc basins were scooped-out by nappes and thrust sheets.

Rifting and wrench faulting of the Himalyan type, which are thought to be the consequence of continent-to-continent collision (Molnar and Tapponier 1975) do not appear to play a major role in Europe. However, modifications in the convergence direction between colliding continents leading to their lateral translation can induce the development of complex wrench and rift systems as for instance during the latest Carboniferous-Early Permian terminal suturing phase of Pangea and during the Neogene development of the Alpine fold belt. Pull-apart features at the termination of wrench-faults such as the Permian Oslo Graben can be highly volcanic.

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ACTIVE RIFTS

During rifting phases the continental crust is stretched and thinned in response to regional extension whereby necking is attained by listric normal faulting at shallow levels and at depth predominantly by ductile flow (McKenzie 1978, Bott 1979, Bally et al. 1981). However, the amount of crustal thinning observed, for instance across the North Sea Rift and the Western Approaches Shelf (Avedik et al. 1982), cannot be fully accounted for by crustal stretching alone. It is therefore inferred that during periods of active crustal extension thermally induced physico-chemical processes affecting the lower crust cause an upward displacement of the curst-mantle boundary and thus also contribute significantly to crustal thinning. These processes, which are vaguely circumscribed as subcrustal erosion, are apparently irreversible. In view of the density differences between the lower crust and the upper mantle subcrustal erosion is another factor that governs the subsidence of a rift zone; moreover, it may function independently from further crustal extension after an initial thermal anomaly has been induced by earlier phases of crustal stretching. At the same time heating of the lower crust increases its ductility so that further crustal extension may not cause deep crustal fracturing facilitating the ascent of magmas to the surface. The Late Jurassic and Early Cretaceous development of the North Sea Rift may be explained by the interaction of the above discussed mechanisms.

Many rifts are totally avolcanic or show only a very low level of volcanism (e.g. Triassic rifts of Western and Central Europe). Other rifts display a high level of volcanic activity right from the onset of crustal extension, as for instance the Rhine Graben or become temporarily volcanic after an initial stage of avolcanic subsidence (e.g. Central North Sea Graben). A high level of volcanism is generally associated with the up-lifting of a wide-radius rift dome that is centered over the axis of the respective rift. Crustal extension resulting from the uparching of such a dome is however, rather small and amounts, for instance to some 200m for a dome with a width of 2000 km and a height of 3 to 4 km (Artemjev and Artyuskov 1971). Uplifting of a rift dome can cause a substantial reversal in the subsidence pattern of a rift (e.g. Central North Sea during Mid-Jurassic) and generally induces extensive erosion, particularly over the rift flanks; on a restricted scale this can contribute to crustal thinning (e.g. Vosges-Black Forest rift dome).

Upwarping of rift domes is apparently caused by the emplacement of low-density, low-velocity upper mantle anomalies at the crust-mantle interface (e.g. Rhein Graben). The development of such anomalies is presumably caused during periods of intensified crustal extension by failure of the lithosphere and/or by thermally induced mantle diapirims (Osmaston 1977, Bott 1976). The physical properties of such upper mantle anomalies, also referred to as 'asthenoliths', 'rift pillows' or 'rift cushions', can be explained by melting processes. Magmas intruding the crust from an asthenolith may eventually reach the surface where they display the alkaline felsic-mafic bimodality that typifies continental rifts (Martin and Piwinski 1972, Burke and Dewey 1973). Moreover, the emplacement of such upper mantle thermal anomalies is likely to induce an acceleration of the subcrustal erosion processes.

Triple jucticns where crustal extension is most intense are the likely places for an early manifestation of rift volcanism (e.g. Rhine Graben and North Sea Rift). However, whilst volcanic activity is not necessarily restricted to the actual rift zone the occurrence of lateral volcanic centres appears to be limited to the area of the rift dome and, by implication, to the confines of the correspondent asthenolith. Asthenoliths are thermally unstable upper mantle anomalies and are resorbed into the mantle upon cooling once crus-

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tal extension has fallen below a certain rate or has ceased altogether. Correspondingly the upper mantle displays beneath thermally stabilized palaeorifts a normal density of about 3.3 and volocities of 8.1 to 8.3 km/sec.

PASSIVE MARGINS AND INACTIVE RIFTS

Following crustal separation the subsidence of the newly-formed 'passive continental margins' is controlled by lithospheric cooling and sedimentary loading of the crust. Similar mechanisms govern the subsidence of rifts that have become inactive (Sleep 1973, 1976, McKenzie 1978, Sclater and Tapscott 1979, Watts and Steckler 1979, Jarvis and McKenzie 1980, Royden et al. 1980, Le Pichon & Sibuet 1981).

The amount of subsidence caused by lithospheric contraction is controlled by the magnitude of the thermal anomaly that was induced during the crustal separation or the crustal stretching stage. It is inferred that maximum thermal anomalies induced by rifting are comparatively smaller. Correspondingly, the post-separation development of a passive continental margin probably reflects the decay of a maximum thermal anomaly whilst the subsidence of inactive rifts is governed by the decay of smaller thermal anomalies. On the other hand the post-rifting development of highly volcanic rifts, that were underlain by an asthenolith during their rifting stage may be associated with the decay of larger thermal anomalies than that of avolcanic rifts. In the post-rifting subsidence pattern of volcanic rifts, the resorption of the asthenolith into the mantle by cooling processes presumably plays a significant role. Moreover erosion of upper crustal rocks over the crest of a rift dome can contribute to crustal thinning and thus will be reflected in the subsidence pattern of an inactive rift.

A further aspect that has to be considered in quantitative subsidence models of rifts is the fact that crustal extension and concomittant subcrustal thinning can take place intermittently over very long periods. In long-lived rifts, thermal anomalies induced by, and associated with, crustal extension can already start to decay during periods of decreased rate of crustal stretching. Thus the thermal anomaly associated with a rift may not be at its maximum when crustal distension terminates altogether and the respective rift becomes inactive. Similarly, late rifting pulses may interrupt and even reverse the lithospheric cooling processes.

This is illustrated by the evolution of for instance the Mesozoic North Sea Rift in which the maximum thermal anomaly was presumably induced during the Early Bajocian uplifting of a rift dome, whilst significant crustal extension persisted into Early Cretaceous times; last, albeit minor, rifting pulse occurred during the Early Paleocene by which time the North Sea Rift had finally become inactive altogether. On the other hand, the development of the West Shetland-Faeroe Rift is characterized by a Late Jurassic thermal surge and a second Late Palaeocene-Early Eocene one (Ziegler 1982).

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RIFT BASINS - A TECTONIC MODEL

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Rift basins are characterized by linear depressions filled by thick sediment layers cut by numerous normal faults. Speculations about the origin of rift basins are generally focused on two processes: (1) initiation by a thermal plume in the mantle (Burke, K. and Dewey, J.F., 1973) and (2) passive formation from extension caused by plate motion (Salveson, J.O., 1978). The thermal plume model requires an initial uplift and the creation of a graben from this uplift. Continued extension beyond the formation of a graben relies on plate motions. The model for passive formation of rift basins relies on plate motion from inception.

Constraints regarding the origin of rifting are contained in the geologic data preserved in the sedimentary record of rift basins. Unfortunately, much of these data are buried deeply within the basin and can only be obtained by drilling. However, several rift basins have been found to have prolific petroleum production and so data from exploration activities is available. These data provide the basis for understanding the tectonics of rift basins. Basins where data are available include the North Sea Basins, Reconcano Basin, Rhine Graben, Red Sea-Gulf of Suez Basins and the Gippsland Basin. These examples show considerable variation in basin size, sediment types, and stage of development. They also show significant similarities. All contain extensive normal faulting. Most have elevated thermal gradients and contain volcanic rocks. Significantly, three stages of sedimentation can be recognized. These stages, pre-rift, rift and post-rift, bracket the tectonic development and so contribute significant information.

The occurrence of pre-rift sediments within the basin is contrary to the theory of large-scale uplift prior to graben formation. In a thermal uplift scenario, the pre-rift sediments should be eroded away. Further, the amount of extension which can be generated by an uplift model is very limited. Five kilometers of extension is calculated for the Rhine Graben (Illies, J.H., 1975). The uplift needed to generate 5 km of extension is in the range of 25 to 35 km. Clearly, a model which requires that much uplift is not viable.

The uplifted shoulders of rifts are cited as evidence for inceptual doming (Lowell, J.D. and Genik, G.J., 1972). However, the uplift of the shoulders are expected from the isostatic relations involved in extensional tectonics. This is illustrated by a simple diagram of isostatically balanced columns of lithosphere (Figure 1). The diagram is based on the assumption that the crust fails under tensional stress by brittle fracture and shearing while the mantle thins like taffy by necking. The graben starts dropping as the extension starts. The pre-rift sediments are faulted down and preserved. The effect of 5 km of extension spread over a segment of crust 50 km wide and 35 km thick is the same as unloading 3 km of material from it. As the upper mantle necks, compensating material rises in a wedge shape out of the asthenosphere. A regional uplift results, tilting the shoulders away from the graben. The extension is clearly the cause of the uplift, not the result.

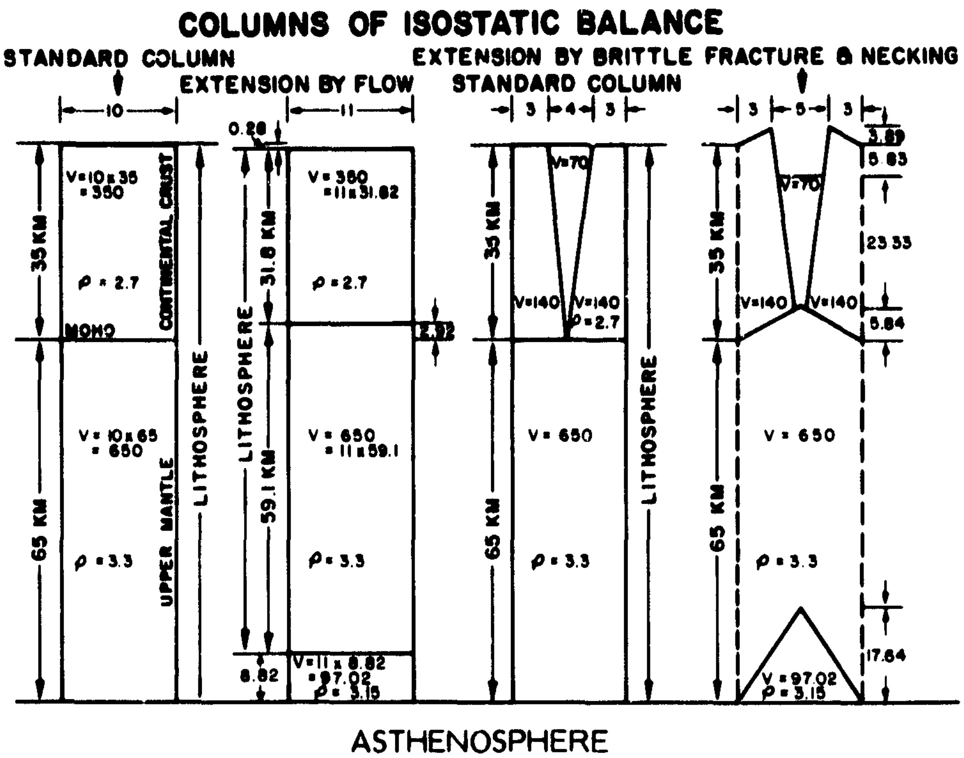

Figure 1.

Standard column at left portrays isostatic equilibrium for 100 km thick lithosphere. Relations required to return to isostatic balance after a 10% extension by flow is shown in next column. Compensating material at bottom of column comes from asthenosphere. Next, another standard column shows V segment where crustal extension is confined for rift model. Last column shows 10% extension where crustal extension is limited to center segment and upper mantle has deformed by necking. Material from asthenosphere flows in from below to compensate for disequilibrium caused by extension. (V is volume of column with unit thickness. ρ is density. The sum of the Vρ products are the same for standard columns but 10% higher for the extended columns).

RIFT BASINS - A TECTONIC MODEL

SALVESON, J. O.

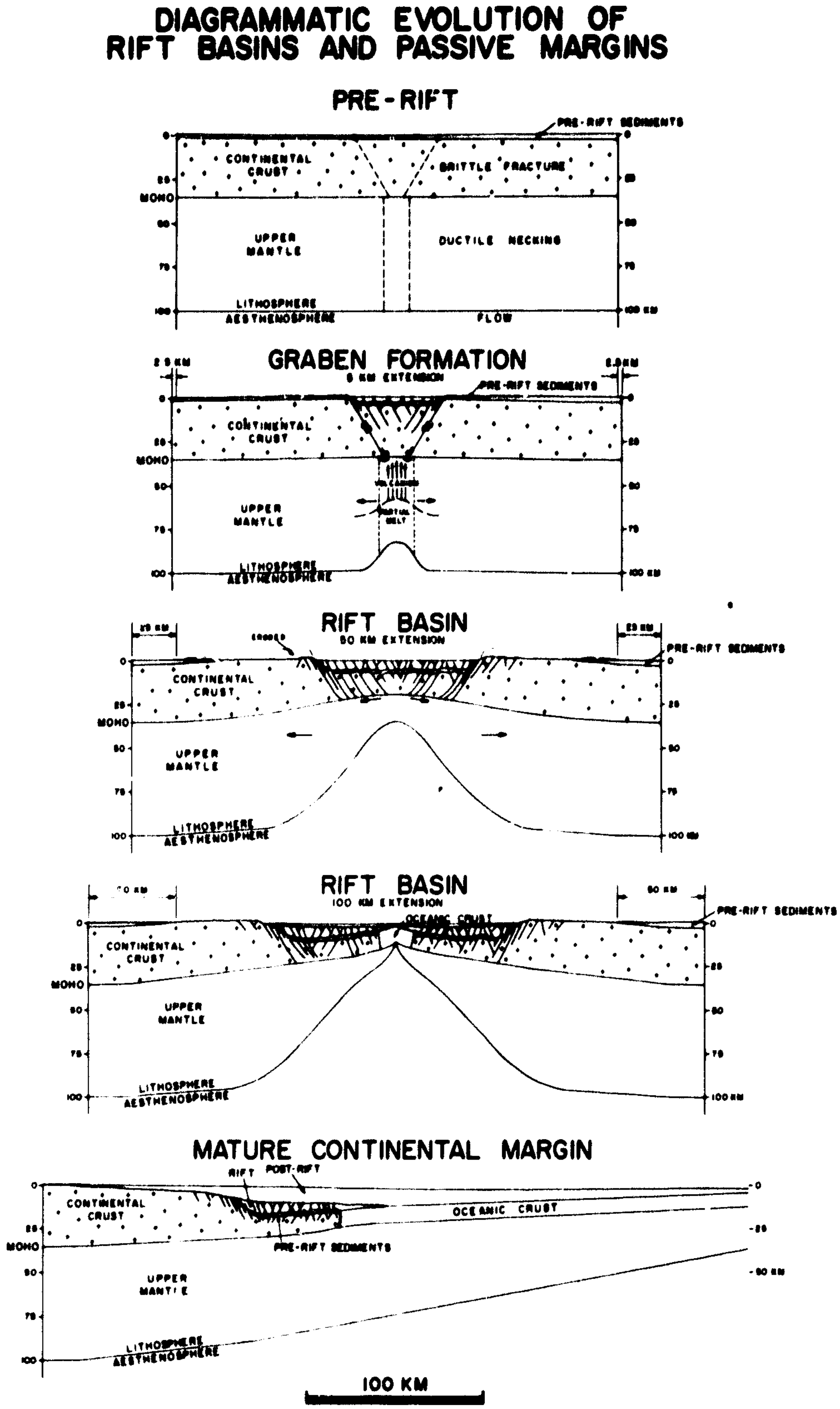

Figure 2

Proposed rift basin model. See text for explanation.

The rising asthenosphere provides a source of increased heat flow. The upper mantle is raised to a decreasing pressure regime and so a partial melt is formed which generates volcanism and adds to the increased heat flow.

An extension rate of 5 mm per year would result in 5 km of extension in only one million years. As extension continues, the crust and upper mantle are thinned until finally the asthenosphere reaches the base of the crust. (See Figure 2). Further extension results in emplacement of oceanic crust and a transfer of the extension to the oceanic spreading center. When faulting of the continental crust abates, the post-rift stage of sediments is deposited. Extension may not always proceed to the formation of a new ocean basin. Any time that extension aborts, faulting stops and the post-rift phase of sedimentation begins. The rift model shown in Figure 2 starts with a segment of lithosphere and proceeds through 5, 50 and 100 km of extension to a mature continental margin. The model is only diagrammatic because every sedimentary basin has unique elements determined by many factors: the thickness of the crust, the effects of previous tectonic regimes, the thermal condition of the lithosphere, the direction and amount of tectonic stress applied, etc. However, the model is scaled and incorporates the isostatic principles discussed earlier. This model of passive response to plate motion allows for considerable variation in the development of rift basins yet explains the common constituents of rift basins: extensive normal faulting, preserved pre-rift sediments, uplifted shoulders, volcanism and high heat flow.

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A HISTORY OF CONTINENTAL RIFTING AT THE MOUTH OF THE GULF OF CALIFORNIA, G. E. Ness and M. W. Lyle, School of Oceanography, Oregon State University, Corvallis, OR 97331

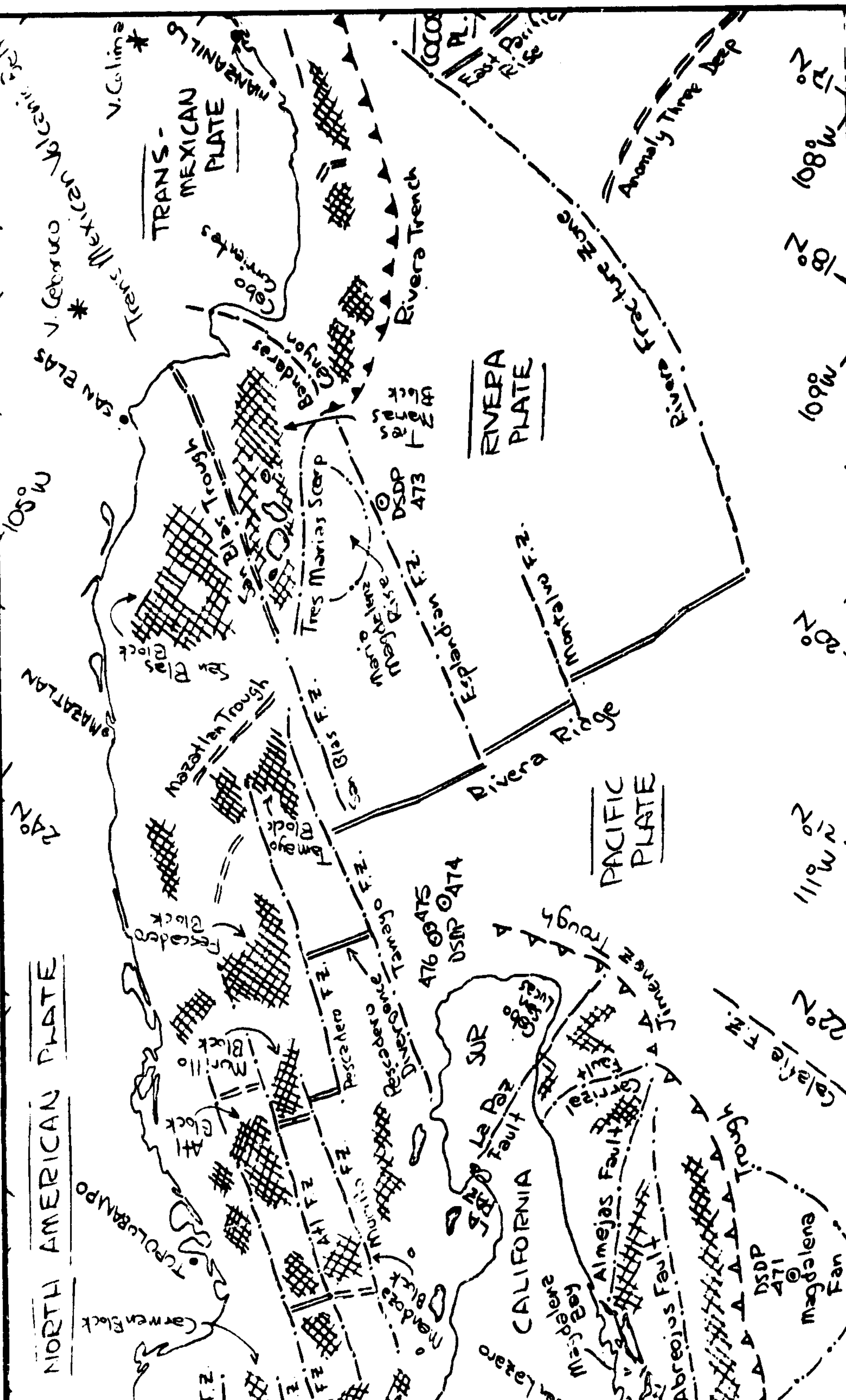
We have constructed a history of the opening of the Gulf of California based upon three major tectonic considerations: that southern Mexico behaves as a separate plate (the Trans-Mexican Plate) from North America, that an episode of subduction has occurred beneath the SE tip of Baja California, and that right-slip motion has occurred on the western margin of the peninsula. These are supported by geophysical and geological evidence.

I. Geophysical Evidence - From magnetic anomaly profiles and newly prepared bathymetric, gravimetric, and seismo-tectonic maps, we identify the active and aborted rifts, the transform faults and their typically buried extensions, and the major continental and oceanic structural blocks in the southern Gulf and the mouth of the Gulf of California. We also identify a buried fossil trench on the west side of Baja California, that extends around the tip of the peninsula, and the geophysical trace of the Tosco-Abreojos fault and two other, still active, faults on the Pacific margin of Baja. Rifting within the Gulf occurred broadly, along multiple divergence zones, leaving foundered blocks in the wake of opening. At least two such blocks are still seismically bounded and independent. Oceanic crust was exposed in the mouth of the Gulf by at least 9 MY based on magnetic anomaly correlations on the northern Rivera Plate. The age of the oceanic crust SE of the tip of Baja is only 3.5 to 4 MY. The Rivera Ridge is uncentered within the mouth of the Gulf as a result of this age difference. This evidence implies that an episode of subduction beneath the SE tip of Baja consumed at least 5 MY of oceanic crust, synchronous with Tosco-Abreojos fault motion. The Pacific Plate is separating from North America at $\sim$70 km/MY, about 25% faster than the $\sim$56 km/MY rate of crustal generation at the Rivera Ridge. This difference can be accommodated by a Trans-Mexican Plate, south of the Trans-Mexican volcanic belt, which moves right laterally to North America at a rate greater than 14 km/MY. This plate boundary passes between San Blas and the Tres Marias Islands.

II. Geological evidence - We have constrained the time and position of opening by postulating that the Magdalena fan (DSDP leg 63 Site 471), a relict feature now situated at the foot of the continental slope SSW of Magdalena Bay (Fig. 1), was deposited at the mouth of the Gulf of California during the initial stages of continental rifting. The quartzo-feldspathic fan sediments were deposited between 14.5 and 13 MY, at a rate of 250 m/MY, which dropped to 50 m/MY after 13 MY (Yeats, 1981). The fan sediments were most probably not derived from the Baja Peninsula immediately to the east because 17-22 MY andesitic volcanics in the region are not represented as clasts in the fan turbidites, and because no source channels or submarine canyons indent the shelf to the NE. The most probable source for the Magdalena fan sediments is the granitic highland around the reconstructed mouth of the Gulf. The Gulf thus began opening about 14 MY, and by 13 MY basins were formed in the interior that trapped the sediments that had fed the fan. If the fan has moved with a Pacific-North American trajectory, it would have occupied a position about 260 km SSE of the present location of Cabo Corrientes at 14 MY, which would require a 19 km/MY rate of NW motion of the Trans-Mexican plate with respect to North America. Gastil and Jensky (1973) argued that a similar offset is needed in the Neogene to match lineaments

GULF OF CALIFORNIA -- RIFTING

Ness, G. E. and Lyle, M. W.



GEOGRAPHIC, STRUCTURAL & TECTONIC FEATURES OF THE MOUTH OF THE GULF OF CALIFORNIA

HACHURES SHOW GRAVITY MAXIMA, TROUGHS ARE GRAVITY OR BATHYMETRIC MINIMA, FAULTS ARE SHOWN IN DASH-DOT LINES, RIDGES AND DIVERGENCES SHOWN AS PAIRED LINES

Figure 1

GULF OF CALIFORNIA -- RIFTING

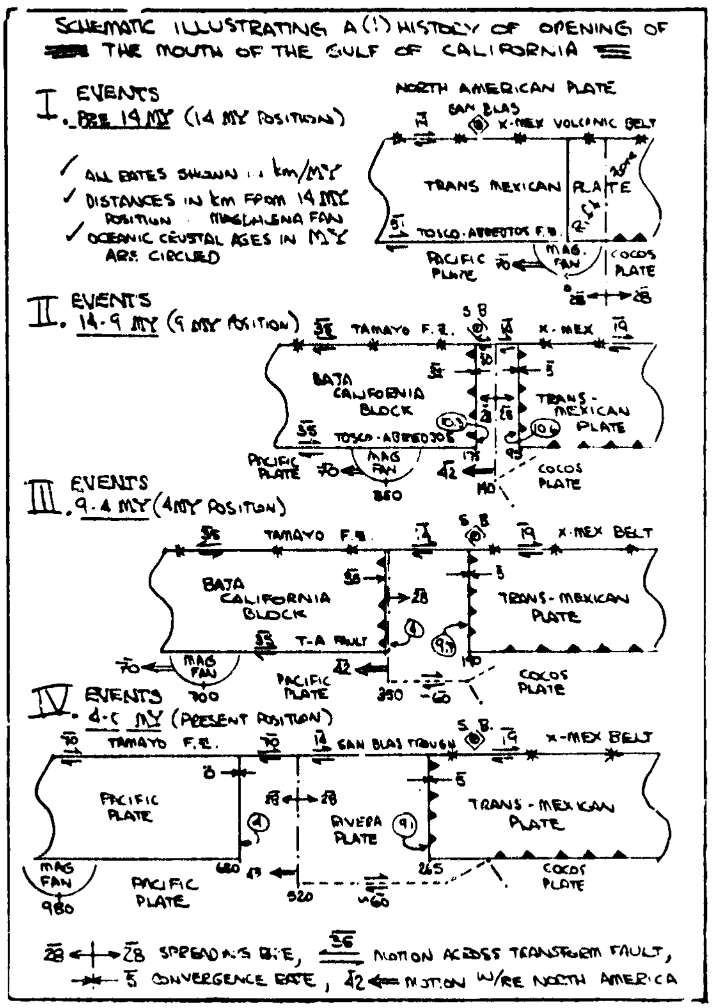
Ness, G. E. and Lyle, M. W.

across the Trans-Mexican volcanic belt. The present position of the Magdalena fan with respect to Baja requires that right slip motion has also occurred along Tosco-Abreojos fault (Normark 1977) probably between 14 MY and 4 MY. Since Pacific crust extends unbroken into the Gulf, subduction must have occurred underneath the SE tip of Baja during the same time interval. Further evidence for subduction comes from the mouth of Gulf transect of DSDP legs 63 and 64. Basement age of Site 474, on oldest oceanic crust at the tip of Baja, is 3.5 MY (based upon microfossils). Oceanic crust formed by the same spreading center on the Rivera Plate (Site 473) has a microfossil age of 6.5 MY. Crust as old as 9 MY may exist at the Rivera Trench, based upon extrapolating the age of the crust at Site 473 and the spreading rate at the Rivera Ridge. Sites 475 and 476, now located at 2650 and 2430 meters respectively on the continental slope at the tip of Baja, help constrain the time when subduction stopped. The two sites have shallow-water sediment sequences in their basal sedimentary sections that have been dated by microfossils (Schrader, in press) and by K-Ar dating of glauconites to have an age of about 4.5 MY. Large scale subsidence of the margin had to have occurred after this time; we postulate that the margin had been dynamically supported by subduction and only subsided after subduction had ceased.

III. Tectonic history - Our reconstruction of the opening of the Gulf is given in figure 2. We have based the reconstruction on the following simplifying assumptions: 1) The initial opening of the mouth of the Gulf occurred at $\sim$14 MY at the backtracked position of the Magdalena fan; 2) PAC/NAM motion has been constant at 70 km/MY along a trajectory that passes through Cabo Corrientes and Cabo San Lucas; 3) The Trans-Mexican Plate has moved with an average 19 km/MY motion to the NW; 4) Neglecting any consideration of its post-10 MY reorientation by $\sim$30°, the Rivera Ridge has had a constant spreading rate of 56 km/MY; 5) Subduction beneath Baja California stopped at 4 MY; 6) The 9 MY position of Baja is simply interpolated between the 14 MY and 4 MY positions which we are more confident of. The PAC/NAM motion is thus equally distributed between the Tamayo fracture zone and the Tosco-Abreojos fault zone over the 10 MY interval; 7) We have adjusted all of our rates to a single small circle extending across the Gulf from Cabo San Lucas to Cabo Corrientes. This simplification does not account for early east-west components of rifting or for possible north-south motion along the La Paz fault. These effects would cause slight rate changes but not affect the general outline of the model

IV. Implications - The terms continental rifting and sea-floor spreading represent slightly different conceptualizations of the same general process of plate divergence. They represent end-member cases occurring in radically different environments, and so they are studied using necessarilly different techniques, by different groups, having different perspectives. The Gulf of California is one of the few areas in the world where continental rifting is just at the point of becoming sea-floor spreading, where, for example, the forms of divergence zones change from basins to ridges. It is, therefore, an opportune place to apply plate tectonics geometrical principles, at the useful mesoscale, to continental structural features. Despite rapid spreading rates, the most striking feature of the history of the opening of the Gulf is the large amount of time involved for the transition to occur. For most of the Gulf the process is still continuing, even after at least 14 MY of activity. The rifting has also occurred over a broad zone, with multiple axes of divergence. Rifting has involved the motion of many small blocks, not just the simple, instantaneous, separation and redirection of a large, monolithic peninsula from the mainland. Rifting has also involved a complicated series of rotations, of kinematic transitions including subduction events. Yet, the record of opening appears to be intelligible to the extent that it may be put into quantitative terms. This is a necessary precondition to understanding not just the kinematics but also the dynamics of rifting.

Figure 2



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RIFTING TECTONICS ON GALICIA PORTUGAL AND N BISCAY MARGIN

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ABSTRACT

The absence of thick sediments on the margins of Galicia, Portugal and North Biscay allows the underlying rift structure to be clearly seen on seismic reflection profiles. The rifted structure consists of a series of tilted, deformed and rotated fault blocks bounded by listric normal faults. The rift fabric has evidently been produced by extension and subsidence.

Listric fault system, extension and thinning of the crust, subsidence during rifting have been interpreted as a consequence of stretching of the lithosphere (MacKenzie, 1978). Stretching process has been quantively examined from subsidence, extension and thinning rates data.

Tilting of the blocks is most probably due to the rotation of the block as it slides down the listric faults. Blocks display also deformation with slight folding of the prerift reflectors, resulting from wedging of the subsiding block between the two adjacent ones. The flattering of normal faults with depth is probably due to an increase of confining pressure and to a slight diminution of the horizontal stress in the direction perpendicular to the rift because of extension.

The extension rate during rifting has been measured from true scale geological sections constructed from migrated multichannel seismic reflection profiles. For a given fault block, the extension rate is by definition the ratio of the horizontal lenght of a block including the lateral offset due to faulting and the length of the pre-rift horizon between the two listric faults bounding the block.

Mean extension rates for the whole margin are 1.10-1.45 and 1.10-1.30 for North Biscay and Galicia Portugal margin respectively. Extension rate may locally reach 1.70 in a single block. In the thinned continental crust adjacent to the ocean crust in Biscay, the extension rate is between 1.10 and 1.40. Thinning rate of the crust at a given point is the ratio between crust thickness near the continent and ocean respectively. It reaches $5^{\pm}1$ near the ocean.

From the present length of the margin of 200 km, an original horizontal length of 162± 20 km is computed from the global extension of the margin. The present section area of the crust of the margin is 3600 km2, and the original area would be 4860± 600 km2 assuming an original crust thickness of 30 km. This mass difference may be explained in three different ways. (1) some mass of the crust was last during rifting, (2) Moho was yet

RIFTING TECTONICS ON PASSIVE MARGINS

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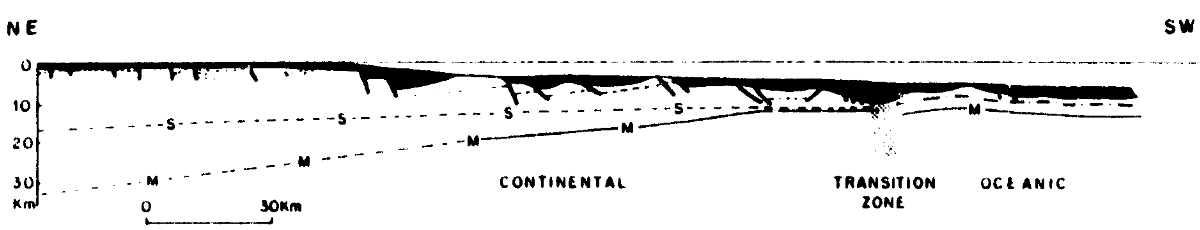
higher than 30 km depth at the beginning of rifting, mass of the crust being conserved during stretching or (3) Moho depth has been modified after rifting.

Refraction data on the N Biscay margin (AVEDIK et al, 1981) show the existence of a strong reflector (S reflector of fig. 1) at between 8 to 10 km depth near the ocean, where listric faults are nearly horizontal. This reflector apparently deepens to 14 ± 1 km toward the continent. It would correspond to the boundary between brittle fracture and ductile flow.

The present section area of the upper brittle crust is 1900 ± 200 km2 (Fig. 1 and 2.a). If it is assumed that brittle-ductile boundary was at 1 ± 2 km depth just before extension, the original section area should have been 2240 ± 250 km2 (Fig. 2.b). To better fit the section area before and after extension if mass conservation is supposed in the brittle crust, it is necessary to imagine a shallower depth of the brittle ductile boundary near the ocean (Fig. 2.c). This depth may be estimated from the total thickness of the layers of the tilted blocks, which is 7 to 8 km. Assuming a 8 km depth for the brittle ductile transition within 60 km near the continent-ocean transition before extension, the area of the brittle crust before extension would be 1850 ± 200 km2 (Fig. 2.c).

This result suggest that some kind stretching of the brittle layer was likely to occur during rifting to account for extension and thinning of the crust but that hot temperatures within the crust just before extension may have provoked shallowing of the brittle ductile-boundary. It is probable that both processes are linked but simple stretching models as proposed by MacKenzie (1978) can't explain the thermal pertubation just before extension.



SCHEMATIC SECTION OF N. BISCAY MARGIN

after AVEDIK et al. 1981

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The question of the mass gap in the lower crust remains open however. The shape of the listric faults as well as the existence of the reflector within the crust that could be considered as a decoupling surface suggest that the lower crust could have undergo additional stretching, but it is highly probable that metamorphism could lead to the observed shallow Moho near the ocean.

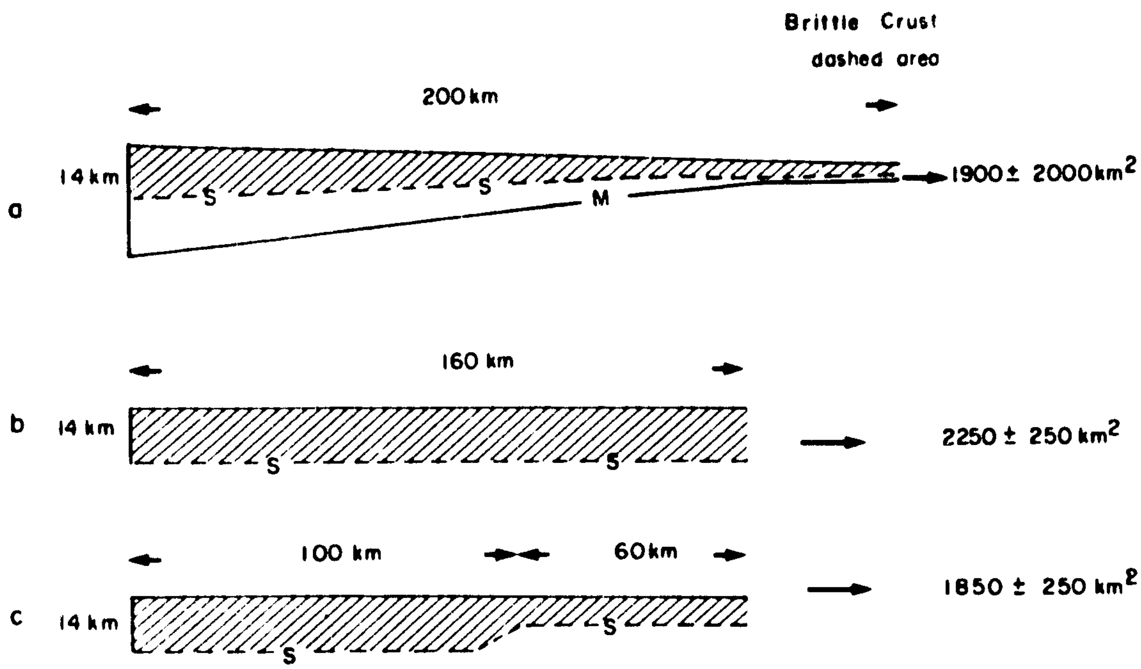

Fig. 2 Stretching of Brittle Crust

RIFTING TECTONICS ON PASSIVE MARGINS

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THE ROLE OF RIFTING IN THE TECTONIC DEVELOPMENT OF THE MIDCONTINENT, U.S.A.; Keller, G.R., Dept. Geol. Sci., Univ. Texas, El Paso, El Paso, TX 79968; Lidiak, E.G., Dept. Earth and Planet. Sci., Univ. Pittsburgh, Pittsburgh, PA 15260; Hinze, W.J. and Braile, L.W., Dept. Geosci., Purdue Univ., West Lafayette, IN 47907

Rifts are widely accepted as important tectonic features of continental margins. However, increasingly they are recognized as a dominant factor in intra-plate tectonism either in their original development or through subsequent control of diastrophism. This is the case in the midcontinent of the United States where rift zones are more common than generally realized. However, paleorifts are difficult to identify because of burial beneath younger deposits, superimposed structures, and subsequent changes in the properties of the underlying crust and upper mantle. Because of these difficulties a review of the geological and geophysical characteristics of the rifts is critical to the definition of those criteria which can be used to identify ancient rifts. Geologically, many rifts are characterized by a complex graben structure in which normal predominates over reverse and strike-slip faulting, by a similar width (35-60 km), by compositionally diverse but generaly bi-modal igneous activity, by transgressive sedimentation except in initial stages, and by an apparent moderate geothermal gradient that results in appreciable metamorphism only in the deeper structural zones. Rift valleys are present only in the younger active structures; older rifts occur mainly as structurally complex zones without primary topographic expression. The rifting process generally involves both the basement and the overlying strata. The geophysical characteristics are equally distinctive. Active rifts are commonly underlain by an anomalously hot crust and mantle. They are charactarized by a thin crust (< 35 km), low Pn velocity (< 8.0 km/sec), long wavelength Bouguer gravity lows on which are superimposed local maxima due to igneous activity and minima due to sedimentary graben fill, shallow Curie temperature depths, complex magnetic anomaly patterns, linear bands of shallow seismicity, extensional earthquake foci, and linear heat flow highs.

Complexities in the geological and geophysical expression of rifts are also present and are due to stage of development at arrestment, age, depth of erosion or burial, and mode of origin. For example, older, paleorifts may be underlain by a more normal, thicker crust and thus lack many of the characteristics of active rifts. Another factor is that many older rifts, particularly those of Precambrian age, differ from younger rifts in being associated with Bouguer gravity highs rather than lows. This difference generally reflects more voluminous basaltic vulcanism or exposure of deeper crustal layers. It may also be important to distinguish between rifts formed by different mechanisms. For example, "dynamic" rifts formed by forces originating from mass transfer within the asthenosphere are expected to have different deep crustal properties than "passive" rifts caused by forces originating within the lithosphere. Vertical movement in the mantle is of prime importance in the development of "dynamic" rifts, and normal faulting and volcanism are manifestations of this larger feature which involves doming and thinning of crust. Such rifts apparently form by one or more of the following mechanisms: development of large thermal anomalies in the mantle, penetration of an oceanic ridge beneath a continent, formation of aulacogens, complete or incipient break-up of a continent, or spreading behind an active continental margin or island arc structure. In contrast, "passive" rifts form primarily as a result of horizontal movement of lithospheric plates. Such rifts may form by the collision of irregular continental margins or by the complex accommodation of microplates to the movement of larger lithospheric plates.

RIFTING IN MIDCONTINENT, U.S.A.

Keller, G.R. et. al.

Although dating of rifts is subject to considerable uncertainty because they are commonly reactivated, a chronological catalog of recognized rifts of the central and eastern midcontinent has been prepared based on a variety of geological and geophysical evidence. The list which includes more than 20 rifts exclusive of late Precambrian and Triassic grabens along the continental margin at time of break-up shows that late Precambrian and Eocambrian rifts are more common, more widely spread and are marked by greater igneous activity than subsequent rifts. Tectonic models developed for the Mississippi Embayment and West Texas areas involve Eocambrian rifting and a subsequent history which has largely been controlled by these ancient features.

RIFT PROPAGATION : THE AFAR CASE

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How does a plate, particularly one carrying continental lithosphere, break ? At what velocity does the rift propagate ? Clearly, the successful theory of plate tectonics, which implies an equilibrium state, cannot answer these important geodynamical questions, which on the contrary imply a transient state. Of related interest is the question of determining the scale at which the plate kinematics model fails and diffuse deformation (which can be constrained by field observations) must be invoked.

Recent progress in this field has been made by Hey and colleagues with particular reference to the Galapagos rift and by Courtillot and colleagues working on the Afar rift zone. Despite similarities in the two approaches, both primarily based on magnetic observations, there remain some important differences. In the Galapagos study (Hey,1977;Hey and Vogt,1977;Hey et al, 1981),the rift propagates along a pre-existing plate boundary, and its forward motion is accompanied by the receding motion of another rift segment. The local kinematics can still be interpreted in terms of plate tectonics (with a continuously moving rotation pole), except in a small area where compression occurs. However, these observations of a propagating/receding pair of rifts (moreover cutting through oceanic lithosphere) is not directly relevant to the question of continental breakup by rifting.

More relevant, in our opinion, to this question is the Afar study (Courtillot and Le Mouël,1978;Courtillot,1980a and b;Courtillot et al,1980). Afar is apparently on the boundary between the Arabian and African plates. Plate kinematics models predict that, in the Afar area, the relative motion must be Northeast-Southwest,at about 1 cm/y., around a rotation pole located thousands of kilometers away. Yet,one can (almost) still walk on dry continental crust from Arabia to Africa, through the Bab-el-Mandeb straights, the Danakil block and the gulf of Zula. Several authors have accounted for this anomaly through the introduction of one or more microplates. The typical (small) size of these microplates and the typical (large) width of their boundaries are such that the plate concept becomes actually meaningless.

The observation of magnetic anomalies over the western gulf of Aden and eastern Afar clearly reveals two distinct magnetic styles. Vine-Matthews like anomalies penetrate as a wedge-shaped zone into a quiet zone: the age of the boundary between the two is found to become younger as one goes West. This is interpreted in terms of the westward propagation of a rift through the lithosphere, at an approximate velocity of 3 cm/yr. The crack tip now lies close to the salted lake Asal, an area which recently suffered seismic and volcanic events (Anis et al,1979). This interpretation is reinforced by observations of bathymetry,seismicity, petrography and geochronology (Courtillot et al,1980).

In this Afar case, there is no evidence whatsoever for a receding rift, such as suggested by Hey and colleagues for the Galapagos, and the area in which plate tectonics theory fails and diffuse deformation occurs is much larger. A major feature in Afar is the occurence of eight axial volcanic ranges (Barberi et al,1975). These are apparently finite-length tears generated in the pre-existing lithosphere by the regional NE-SW tension. A scenario of how the rift propagates can be summarised as follows (Courtillot, 1980 b) : in the initial phase, the pre-existing lithosphere undergoes considerable mechanical and thermal modifications by normal faulting, dyke injection (as currently observed in Afar) and some amount of anelastic defor-

RIFT PROPAGATION

Courtillot, V.

mation. This generates the magnetic quiet zone. Tensile stress ahead of the rift tip is locally relaxed by the generation of axial ranges: they are associated with pseudo-oceanic magnetic anomalies, which can hardly be correlated with reversal time scales (e.g. the Manda-Inakir range). Several such tears can be formed perpendicular to the regional principal tensile stress direction. As the main rift propagates, the relative positions of the ranges and the rift tip changes and a tectonic and chemical evolution of the ranges occurs. Some ranges become extinct, new ones may form. The propagating rift may preferentially use a pre-existing volcanic range along its way. Normal oceanic crust, bearing typical magnetic anomalies, is generated in the wake of the rift tip, with the observed wedge-shape. The future evolution probably involves the NW propagation of the rift towards the Red Sea trough (a similar South propagating rift ?).

The model still raises a number of important mechanical questions; several of its predictions can be tested. In order to obtain more data relevant to this problem, a joint paleomagnetic, tectonic and geochronologic study was performed in Afar in the Spring of 1981. 40 sites were sampled, South of lake Asal, in an area 50 by 100 km. The study team consisted of :

. paleomagnetism : V. Courtillot and J. Achache (Univ. of Paris)
N. Bonhommet and P. Galibert (Univ. of Rennes)
. tectonics : P. Tapponnier (Univ. of Paris)
F. Arthaud and P. Grelet (Univ. of Montpellier)
. K-Ar dating : G. Féraud (Univ. of Nice)
R. Montigny (Univ. of Strasbourg)

It is expected that paleomagnetic measurements, coupled with a detailed tectonic study and age determinations, will allow one to reconstruct the history of deformation as the rift propagated from the gulf of Aden 4 Myr ago to its present location. The study of declination should allow one to map the horizontal rotations and that of inclination to study the history of acquisition of tilt, another interesting and partly unanswered question in the evolution of a spreading center (see f.i. Tapponnier and Francheteau, 1978). Preliminary results, as available, will be presented at the conference.

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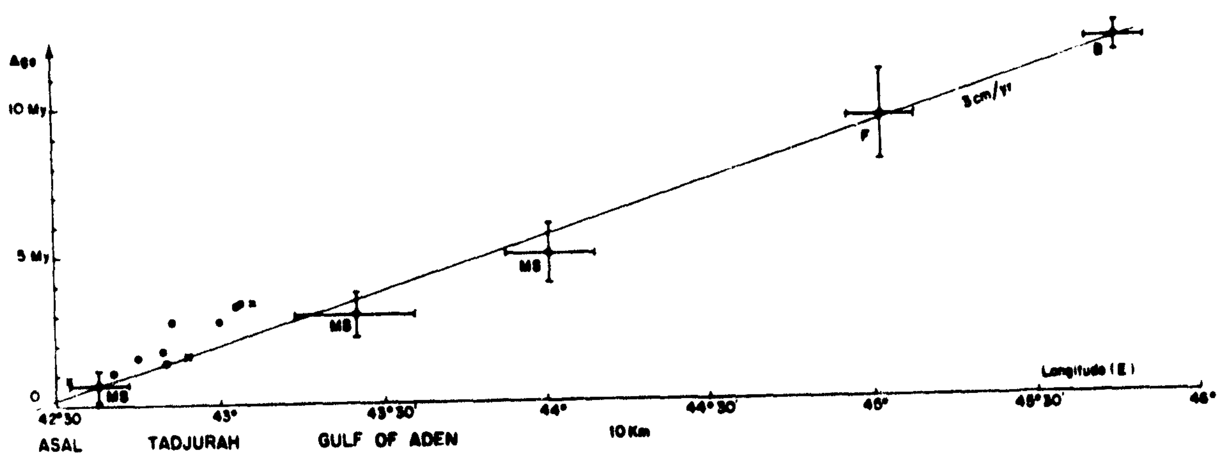

Figure (from Courtillot et al,1980) : age of the boundary between the quiet and oceanic zones using various data sources. This figure is the basis for the evaluation of the propagation velocity of the rift.

(Figure published courtesy of Elsevier Scientific Publishing Company)

CONTINENTAL RIFT JUMPS? C. A. Wood, SN6/NASA Johnson Space Center, Houston, TX 77058

Detailed studies of magnetic and bathymetric records suggest that oceanic spreading ridges sometimes discontinuously shift to a nearby, often parallel location.[1,2,3] Such jumps are thought to return a migrating ridge axis back to the hot spot that provides energy for sea floor spreading. Roughly parallel trending patterns of faults and volcanoes along various rift valleys suggest that similar discontinuous "rift jumps" occur in continental terrains. In this note possible rift jumps in Iceland, West Africa, and East Africa are described, leading to speculations on the mechanism of such jumps.

Iceland Jumps. Iceland provides a fitting transition from ridge jumps to rift jumps in that it is physically and compositionally intermediate between oceans and continents. Although the Icelandic rift system is a direct continuation of the oceanic ridges to its north and south, it appears to be more complex, having 2-6 active, subparallel rift zones, instead of the single spreading center of the ocean ridges.[4] Abandoned rift zones (Fig. 1) are identified in eastern (extending south from the Holaheidi peninsula) and west-central (from Skagi peninsula south to the current rift zone at Thingvellir) Iceland,[4,5] and it is possible that each of the penisulas on Iceland's northern and western shores represents an ancinet rift. Each of the identified rift zones has a relatively short period of activity, and some, such as the closely spaced and parallel Grimsvofn and Torfajokull zones,[4] are active contemporaneously. Walker did not speculate on why new rift zones have been generated so frequently in Iceland, but Saemundsson[6] proposed that the major spreading jump from western to eastern Iceland, which occurred ~7-8 m.y. ago, was due to an eastward migration of an underlying mantle plume.

Afar Jumps. The Afar region of Ethiopia, an area of continental crust attenuation and ocean crust formation, includes a number of proposed spreading ridge jumps from the Red Sea into Afar, providing another intermediate step from ridge to rift jumping. Barberi and Varet[7] have suggested that the Erta Ale range and other lines of tholeiitic volcanoes in northern Afar are spreading ridges connected to each other and to the Red Sea by transform faults (Fig. 2). A remarkable feature of this proposal is that the main jump into Afar isolated a large continental massif - the Danakil Alps - from the main African plate. Additionally posited in Afar are a number of paired spreading ridges, somewhat similar to the coupled rift zones in Iceland but with wider spacings (~40 km vs. 10-20 km in Iceland).

Benue Trough-Cameroon Volcanics. The approximate parallelism of the Benue Trough and Cameroon Volcanic Line and their remarkable similarity in shape have led Fitton[8] to propose that both are related to a common asthenospheric hot zone under West Africa. The Benue Trough is interpreted as a failed rift arm associated with the Cretaceous opening of the South Atlantic,[9] and the Cameroon Line is a 1600 km long trace of alkaline volcanics that have been active since 65 m.y. ago.[8] Fitton suggests that a 7° clockwise rotation of Africa ~80 m.y. ago shifted the Y-shaped asthenospheric hot zone from the Benue Trough to the position of the subsequent Cameroon Volcanic Line.

Kenya Rift Jumps. Faulting, uplift, and volcanism have occurred in Kenya since the early Miocene. Both faulting and volcanism (Fig. 3) have systematically shifted eastward with time.[10,12] The earliest recognizable faults are monoclinal flexures in eastern Uganda[10] along which occurred Miocene central volcanoes of nephelinitic affinity. Miocene age basalts erupted in the trough where the rift valley later developed, and nearly all Pliocene volcanism (mostly phonolitic and trachytic) was centered in the rift.[12] By 5 m.y. ago

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basalts had appeared east of the rift,[12,13] and most of the Quaternary[13] basalts in Kenya occur well to the rift's east. The large central volcanoes, Kenya (2-3.5 m.y. old),[12] Kilimanjaro (0.4-1 m.y.),[12] and Marsabit (Quaternary),[13] also formed east of the rift.

Rift faulting followed a similar but less well dated progression eastward (Fig. 4), with early and middle Pliocene activity centered in western Kenya and late Pliocene to Quaternary faulting along the main rift axis.[10] The migration of rifting eastward is especially well displayed in northern Kenya.[14] There is little evidence for faulting east of the rift, however, except for the alignment of volcanic centers.

Volcanism east of the rift includes a number of maar craters which formed in areas where near surface water probably was not available to cause their explosive activity - e.g., El Sod in southern Ethiopia, and the crater fields of Marsabit, Leisamis, and Nyambeni in northern Kenya. Non-phreatomagmatic maars, such as these appear to be, are thought to be surface expressions of deep seated diatremes,[15] and indeed, ejecta of El Sod include perioditite and ultramafic granite.[16] I propose that these maars and associated basalt fields are places where the crust has been fractured, and mark the location of future rift development.

Sparse geophysical evidence lends some support to this proposal. Models of gravity data[10] indicate that the anomalous asthenosphere extends 20-30 km east of the rift axis, but only under the rift do massive intrusions fully penetrate the crust. Geomagnetic deep sounding also indicates regions of high electrical conductivity (partial melts) east of the rift and perhaps 100 km deep.[17]

Discussion. The examples discussed above demonstrate that continental rifts shift laterally to new locations, a process perhaps grossly similar to ocean ridge jumps. In each case discussed the shift is of a discrete distance, not a gradual migration; each rift zone is cleanly separated from its neighbor. The jump distance increases from ~10-20 km in Iceland, to ~40 km in Afar, to 100-175 km in Kenya, and to ~250 km for the Benue-Cameroon jump. These values are approximately equal to lithosphere thickness in each of these areas (Fig. 5), suggesting some mechanism of lithospheric control on jumping. Volcano-tectonic activity is not synchronous along parallel rifts; in Iceland volcanic activity continues at the souther ends of various rift zones whose northern ends are mostly extinct.[4] Similarly, volcanism in the three postulated Kenya rifts began in the north and spread southward. For both the as yet unformed east Kenya rift and the Cameroon volcanism there is abundant volcanism along a well defined line, but no significant rift faulting. Thus, discrete patches of volcanism develop before any independent evidence of tectonism.

Rift jumping must represent a displacement between underlying magma sources and the crust/lithosphere. A slight rotation of Africa may explain the Benue-Cameroon jump,[8] but there is no evidence that Africa lurched westward ~12 m.y. ago and again ~10 m.y. later, to account for the episodic development of the Kenya rifts. It is more likely that anomalous asthenosphere currently beneath the main Kenya rift is moving eastward under Kenya, being pulled by the eastward movement of Arabia and the Red Sea spreading system.[18]

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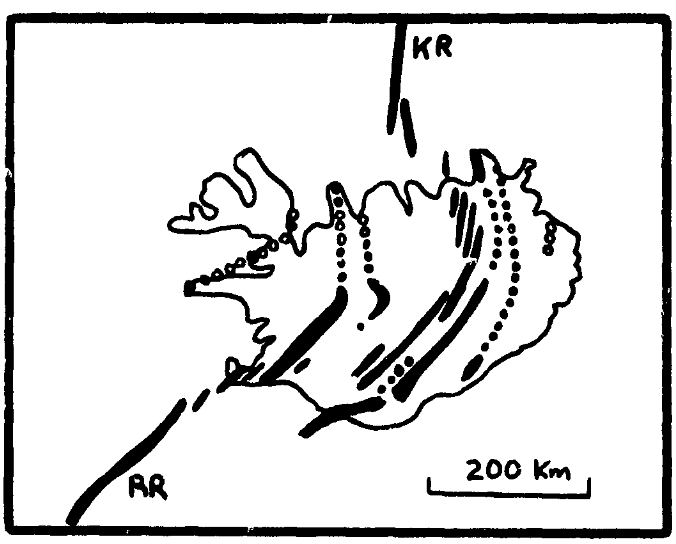

Fig. 1: Schematic chart of spreading ridges in Iceland, after ref. 4.

Active Rift Zone (RZ)

● Dormant RZ ○ Extinct RZ

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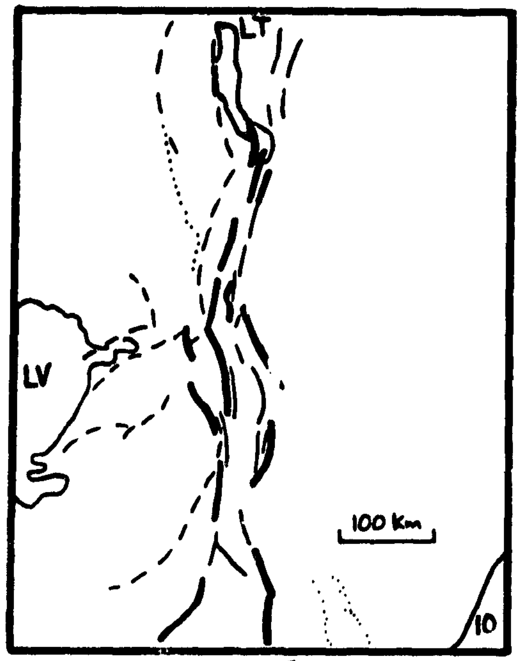
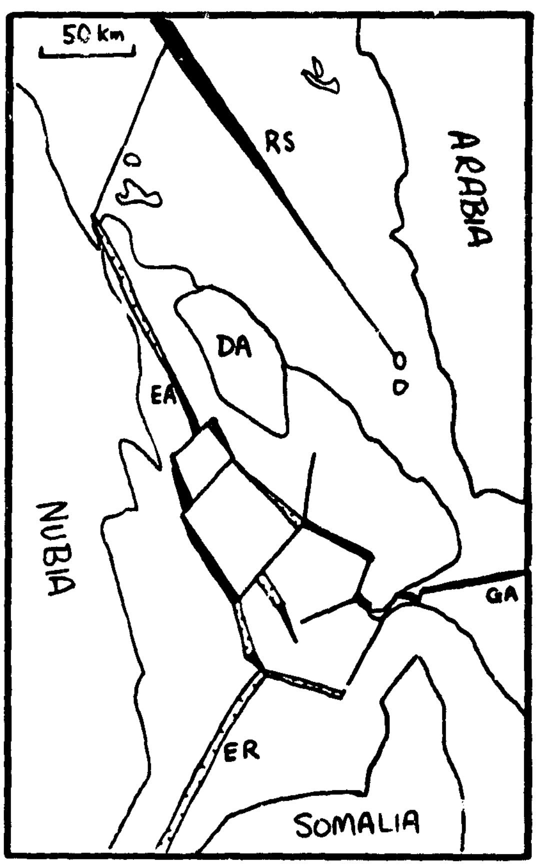
Miocene Nephelites   Miocene Basalts

Pliocene Basalts   Quaternary Basalts



Miocene   L/M Pliocene

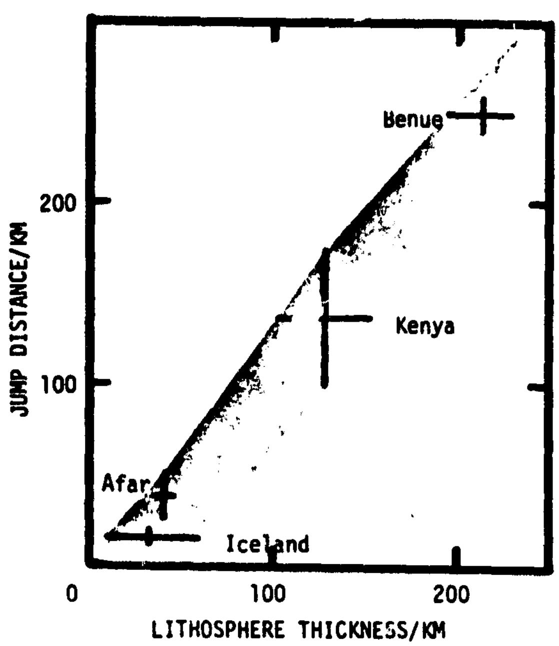
U Pliocene + L Pleistocene   L Quaternary



Spreading Ridge   Extension Axis

Transform Fault

Fig. 2 (left): Spreading ridges in Afar, from ref. 7. DA=Danakil Alps, EA=Erta Ale, ET=Eth. Rift, GA=Gulf of Aden, RS=Red Sea. Fig. 3 & 4 (above): Volcanism and faulting in Kenya, ref. 10. IO= Indian Ocean, LT=Lake Turkana, LV= Lake Victoria. Fig. 5 (below): Lithosphere thickness (various sources) and rift jump distance.



CONTINENTAL AND OCEANIC RIFT/TRANSFORM SETTINGS: SIMILARITIES, DIFFERENCES AND RELATION TO LITHOSPHERIC EVOLUTION

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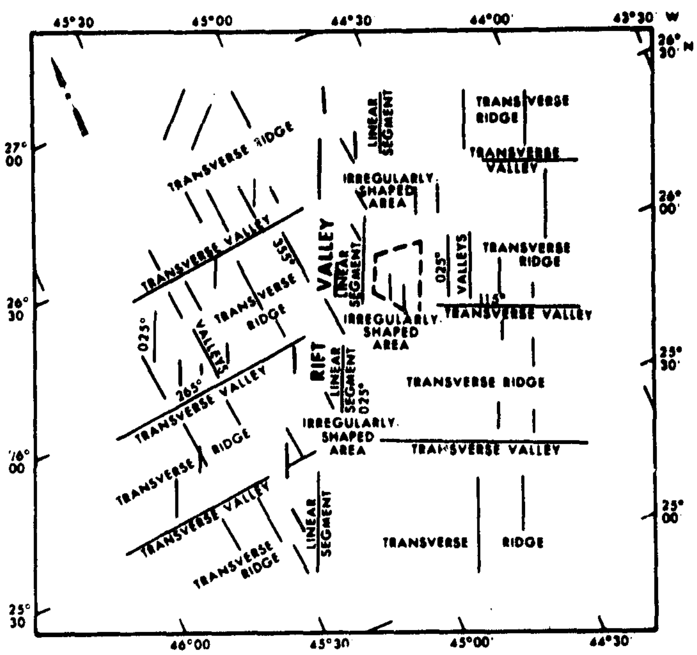
Oceanic and intracontinental rifts and transforms exhibit similar physiographic features, types of igneous activity, and geothermal characteristics. For example, transverse ridges and valleys (figure), sedimentary basins at rift-transform intersections, spacing between transforms, and variability in lengths of transforms are surprisingly similar for the Kane fracture zone (Rona and others, 1976), the FAMOUS area (Phillips and Fleming, 1978), and the southern Rio Grande rift (Dickerson, 1980).



(Rona and others, 1976)

(Figure published courtesy of the Geological Society of America)

The manifestations of these features in gravity and magnetic data are similar as well: gravity data for the Oslo rift (Ramberg, 1976), those for part of the Rio Grande rift in New Mexico (Ramberg and others, 1978), for the Gulf of Tadjurah (Courtillot and others, 1980), for the FAMOUS area (Phillips and Fleming, 1978), and for the Kane fracture zone (Rona and others, 1976).

The magnetic data presented by Courtillot and others (1980) for the Gulf of Tadjurah, however, have great relevance in any comparison of oceanic and continental rifts and transforms. Their survey area comprises on- and offshore portions of a propagating rift zone and related transforms. For one transform in particular, there is a good correlation on land between a strong positive anomaly and outcrops of partly submarine basalts; the same basalts, also associated with a strong positive anomaly, are found offshore and are interpreted as representing older oceanic sea floor offset by a transform fault.

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Importantly, the character of the whole suite of those anomalies varies little from onshore to offshore.

Hydrothermal activity and higher heat flow values in rift zones are well documented; greatest activity and highest heat flow values are commonly at offsets of the rift axis, as in the Rio Grande rift (Reiter and others, 1978) and the TAG hydrothermal field of the Mid-Atlantic Ridge (Rona and others, 1976). Numerous hot springs and occurrences of hot waters in wells have been mapped along the bounding faults of the southern Rio Grande rift; there, too, the greatest concentrations coincide with offsets in the rift axis, as at the intersection of the rift and the Shafter zone (Dickerson, in press). Significantly, hot springs are also common along the Shafter zone itself, which I interpret as a boundary transform to the rift.

Evolution of laterally and vertically stable continental lithosphere about 1.9 b.y. ago may have been accomplished, in part, through rifting and emplacement of basalts now preserved as greenstone belts. The rhyolite-basalt association common in Cenozoic rifts is also common in such early Precambrian greenstone belts; the older greenstone belt volcanism was similar to that now producing tholeiites in modern oceanic and continental rift zones. Although no basins comparable with modern oceans were formed during early Precambrian rifting, the number of greenstone belts preserved in Precambrian shields is so great that the total width of those sequences is close to that of the Atlantic Ocean, according to Grachev and Fedorovsky (1981).

Continental splitting in response to mantle convection could begin when the strengths of the oceanic and continental types of lithosphere became sufficiently close (Tarling, 1980). There is increasing evidence for ancient rift zones, among which is the structural and lithologic information from greenstone belts in the shields. If ancient transforms were also operative, criteria for recognizing them should include magnetic and gravity expression, presence of pods of metasediments at junctions of rifts and transforms, offset metabasalt or greenstone belts, and distribution of mineralized zones of hydrothermal origin.

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SURFACE DISPLACEMENT DURING DIKE INTRUSION AND FISSURE ERUPTION IN VOLCANIC RIFT ZONES. D. D. Pollard and W. A. Duffield, U.S. Geological Survey, Menlo Park, CA 94025.

Magmatic activity within the rift zones of Kilauea Volcano is characterized by shallow intrusive events and fissure eruptions (Swanson et. al., 1976). Recognizing that a great variety of geological, geochemical, and geophysical observations are pertinent to a complete understanding of this activity, we focus on deformation at the Earth's surface as indicated by relative displacements. Our principle objective is to develop a method for determining the geometric parameters (height, depth, and inclination) and mechanical parameters (magma driving pressure and host rock stiffness) of the source of this deformation.

Deformation studies of the summit region of Kilauea Volcano (Fiske and Kinoshita, 1969) have identified a possible reservoir of magma that systematically inflates and deflates. Summit deflation is clearly correlated with intrusion of magma from the reservoir, both upward to the caldera and laterally into the rift zones. A variety of seismic and deformation studies (Koyanagi et al., 1972; Dieterich and Decker, 1975) have suggested that the reservoir is several kilometers deep and approximates a vertical cylinder with a height that is several times greater than the radius. The most comprehensive modeling of the seismicity to this date by Ryan et al. (1981) has revealed a very complex magma transport and storage structure beneath the summit. Individual magmatic zones within this structure may be composed of a plexus of dikes, sills, and more irregular intrusive forms, but more specific details have not emerged.

In contrast, geological observations of eroded rifts (MacDonald and Abbott, 1970), geophysical surveys (Hill, 1969), deformation studies (Duffield, et. al., 1974), theoretical and comparative studies of ancient dikes (Delaney and Pollard, 1981a) and the very nature of fissure eruptions demonstrate that the magma released from the summit reservoir into the rifts travels in nearly vertical dikes. This information on source shape and inclination greatly constrains the scope of our analysis and enables us to develop a simple and very specific method to interpret deformation data in volcanic rift zones.

Analyses of eroded dikes (Pollard and Muller, 1976; Delaney and Pollard, 1981a) show that the dilational form of these intrusions is closely approximated by that of planar cracks in an elastic material subjected to certain distributions of internal pressure and remote stress. To analyze surface deformation due to dike emplacement we adopt the methods developed by Pollard and Holzhausen (1979) for cracks interacting with a stress-free surface. Although this method of solution is restricted to isotropic and homogeneous elastic properties and two-dimensional geometry, it admits any number of cracks of arbitrary length, inclination, and depth subject to arbitrary loads. The cracks may even intersect the free surface to mimic a fissure eruption. Operator convenience, computational efficiency, and accuracy are greatly improved over finite element methods because only the boundaries of the cracks are discretized, and superimposed analytical solutions determine deformation elsewhere.

Fissure eruptions generally last less than a few days before magma solidifies in much of the dike (Delaney and Pollard, 1981b). Unless major tectonic events such as earthquakes, gravity slides, or nearby intrusions occur during this brief time, it is reasonable to assume that remote stresses change very little compared to the changes in internal pressure.

SURFACE DISPLACEMENTS IN RIFTS

Pollard, D. D. and Duffield, W. A.

This enables us to limit attention to displacement fields that originate from internal loading of a crack. The crack dilates because of a driving pressure that is equal to the difference between the magma pressure and the remote least compressive stress. Dilation is resisted by the stiffness of the host rock.

Considerable insight that is valuable when interpreting surface displacement data in rifts may be gained by carefully examining the nature of the source. Because of its geometry, the crack elicits a strongly anisotropic deformation. That is, displacements acting normal to the crack are directed outward away from the crack plane, but displacements acting parallel to the crack are directed inward toward the crack center. This is in marked contrast to more equidimensional sources (a circular hole is the extreme example) which displace outward in all directions when pressurized. This simple fact accounts for a unique bimodal distibution of surface displacement that distinguishes vertical cracks from other sources (see e.g. Dieterich and Decker, 1975, Figs. 3-8). Immediately over the crack the surface displaces downward in keeping with the inward displacement mentioned above. To either side of this trough the surface displaces upward into two ridges to compensate for the outward displacements mentioned above. Relative heights of the two ridges depend upon inclination of the crack. The three geometeric and two mechanical parameters of the source may be extracted from analysis of one vertical displacement profile. Intrusive and eruptive events of May 15-16, 1970 and Sept. 24-29, 1971 on Kilauea are used to illustrate our method of analysis and interpretation.

We believe that this method has potential applications for other Hawaiian events and perhaps in similar geological settings such as Iceland (Sigurdsson, 1979) and East Africa (Tarantola, et. al., 1979). The vertical displacements recorded during a rifting event in Jan. 1978 near Krafla Volcano in North Iceland are used to demonstrate such an application. This event produced vertical displacements consistent with a dike ~3.5 km high witn a dip >85° and a depth-to-center of ~3 km. Given an elastic shear modulus of 4×10^4 MN/m^2 and a Poisson's Ratio of 0.25, we estimate a driving pressure of ~1 MN/m^2. This is sufficient to dilate the dike to a thickness of ~2.5m.

The complete data set on surface displacements in rift zones has rapidly expanded in the past decade as interest has focused on monitoring volcanoes by geodetic means. The method of analysis presented here provides a practical tool that could be used in routine monitoring of rift zones. The physical parameters determined by studies of this kind should improve the accuracy of volcanological forecasting as well as our general understanding of how a rift might develop. For example, one dike emplacement scheme involves great lateral propagation with concurrent increase in height from a fixed depth-to-center (Fiske and Jackson, 1972). A fundamentally different scheme involves a decreasing depth-to-center due to upward propagation of the entire dike, which maintains a fixed height (Weertman, 1971). Our results clearly distinguish these schemes by showing how the surface displacement distributions evolve with time during an intrusive event. The data required for this analysis must include several displacement profiles taken during one event.

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NATURE AND EVOLUTION OF MANTLE SOURCES DURING BACK-ARC RIFTING

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Many rifts on Earth are associated with zones of potential plate separation and lithosphere generation. However another whole series of rifts, back-arc rifts, are connected with plate collisions and lithosphere consumption. Such rifts occur both at continental margins and in intraoceanic areas and with further development give rise to marginal back-arc basins in various forms. This paper concentrates on the geochemistry of lavas erupted during the initial stages of back-arc rifting and attempts to infer from this the nature and evolution of mantle sources underlying the rift zone. Comparisons are made between intraoceanic rifts such as the East Scotia Sea and Western Pacific, continental margin rifts such as Bransfield Strait, Antarctica, southern Chile and the Gulf of California and mid-continental rifts such as East Africa. Archean greenstone belts represent major crustal rifting features formed at an early stage of development of the Earth's continental crust, but seem to have more in common with back-arc rifts than modern mid-continental rift features.

The Scotia Sea, bounded by the extended loop of the Scotia Arc, represents a complex zone of microplates at the boundary of two major plates, the South American and the Antarctic. In the East Scotia Sea back-arc spreading has been underway for about 8 Ma behind the South Sandwich volcanic arc, although magnetic lineation patterns indicate that the arc itself may have been built on lithosphere generated at the spreading center and that spreading is asymmetric. The arc is made up of lavas belonging to the island arc tholeiite series. Geochemical studies of basalts dredged from the back-arc spreading center and from older portions of the back-arc lithosphere (Saunders and Tarney, 1979) show that while most resemble mid-ocean ridge basalts (MORB) some lavas are unusually vesicular, have glassy rinds with higher H_2O contents and higher H_2O/CO_2 ratios than MORB, and have higher K and Rb contents and higher $^{87}Sr/^{86}Sr$ ratios too. The data are best explained in terms of minor contamination of the mantle source feeding the back-arc rift zone with fluids carrying mobile chemical components from the subducting slab (Tarney et al. 1981).

The Mariana Arc system in the Western Pacific is an example of a multiple arc - marginal basin system that has developed since the Eocene following the change in direction of Pacific Plate motion about 40 Ma ago. The initial Palau-Kyushu arc evolved rapidly above a new subduction zone initiated at a major N-S transform fault (Hilde et al. 1977), and the lavas are mainly island arc-tholeiites, but with boninitic lavas erupted on the trench side of the arc. About 10 Ma after the start of subduction, the arc was rifted apart and the two halves progressively separated by the formation of the Shikoku and Parece-Vela Basins by back arc spreading. The new frontal arc remained active, but erupted lavas with more calc-alkaline characteristics. About 6 Ma ago this second arc was rifted apart to form the Mariana Trough and leaving a remnant arc (West Mariana Ridge) and the presently active Mariana Arc. Arc volcanism at each stage shows a chemical evolution from island arc tholeiite toward calc-alkaline, but with the latter characteristic becoming dominant with time overall (Tarney et al. 1981). This is expressed chemically by higher concentrations of elements such as K, Rb, Ba, Sr and the light rare-earths. Interestingly, drilled basalts recovered from the Mariana Trough are vesicular and have trace element characteristics transitional towards calc-alkaline, probably resulting from chemical modification of the mantle source by fluids, as in the case of the Scotia Sea basalts (Tarney et al. 1981).

Such chemical characteristics appear to be dominant during the initial stages of back-arc rifting and particularly where the arc being split has been erupting calc-alkaline lavas. A specially good example of this is the young marginal basin of Bransfield Strait bordering the northern Antarctic Peninsula. Rifting and ensuing back-arc spreading began less than 1 Ma ago, separating the South Shetland volcanic arc from the Peninsular proper. The region has had a continual history of calc-alkaline volcanism since the early Mesozoic. The volcanoes of Deception, Bridgman and Penguin Islands, lying on or close to the axis of back-arc spreading are erupting lavas with transitional ocean-floor basalt - calc-alkaline basalt major and trace element characteristics. (Weaver et al. 1979). The lavas of Penguin Island are in fact mildly alkaline, but unlike alkalic lavas associated with mid-continent rifting are markedly deficient in high field strength elements such as Nb and Ta - a calc-alkaline characteristic.

Similar features are apparent in the Rocas Verdes autochthonous marginal basin ophiolite (Mesozoic age) in southern Chile (Dalziel et al. 1974; Saunders et al. 1979). In the south at Tortuga where the basin was wide, the basalts forming the central part of the basin are essentially similar to MORB whereas those flanking the basin, presumably erupted earlier, are enriched in light rare earths and other incompatible elements. Where the basin narrows to the north, as at Sarmiento (Saunders et al. 1979), the basalts are mostly all of the latter type, and are deficient in Ta and Nb.

The Gulf of California represents a continental margin rift broadly similar to Bransfield Strait and the southern Chile example. Recent drilling on DSDP Legs 64 and 65 has recovered basalts from beneath the thick sedimentary cover which are representative of those erupted during the initial phase of rifting and those erupted when spreading had been underway for some time. Here again there is a transition in geochemistry from those mildly enriched in incompatible elements to typical MORB with time. The trace element characteristics are however more calc-alkaline than alkaline (e.g. Saunders et al, 1982; Saunders 1982).

In summary, most examples of back-arc rifting occur through splitting of an active pre-existing volcanic arc. As with mid-continent rifting the earlier lavas erupted are relatively enriched in incompatible elements, but whereas the mid-continent rift lavas are distinctly alkaline and enriched in Ta and Nb, the continental margin or intraoceanic equivalents are poor in Ta and Nb and have transitional calc-alkaline characteristics (Tarney et al. 1981). Especially where the pre-existing volcanic arc itself has been erupting calc-alkaline rather than island arc tholeiite lavas. The ultimate cause of back-arc rifting may depend on global and local plate interactions, but the focus of initial rifting is strongly constrained at the volcanic arc. Back-arc spreading may occur to a variable extent, but never seems to extend for more than 10-15 Ma before either ceasing completely or jumping to form a new rift beneath the frontal arc.

Precambrian greenstone belts have many features in common with modern marginal basins (Tarney and Windley, 1981) and represent major rift-basin structures which have developed penecontemporaneously with crustal generation.

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Though dominant in the Archean they also extend into the Proterozoic. The differences in form between these and modern mid-continent or back-arc rifts may reflect differences in the strength and maturity of the underlying lithosphere.

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RIFTING WITHIN THE AFRICAN CONTINENT: A MODEL FOR CONTINENTAL DISRUPTION? J. D. Fairhead and S. E. Browne, Dept. of Earth Sciences, University of Leeds, Leeds LS2 9JT, UK

This contribution investigates whether the East African Rift System or the hitherto little known subsiding rift system of Central Africa are typical models for continental disruption.

The best example of intraplate volcanism is that associated with the East African Rift System where extensive geophysical evidence indicates that the crust is underlain by an anomalously hot, low density, low seismic velocity body at shallow depth in the mantle (Gass et al., 1978). This body is widely thought to represent, in plate tectonic model terms, asthenospheric material at shallow depth beneath a lithosphere which has been thinned by the upward migration (stoping) of the lithosphere-asthenosphere boundary within the last 45 Ma. The domal uplift regions of Ethiopia and Kenya are further considered to be sites within the Rift System where the lithosphere has undergone the greatest amount of thinning and consequently represents the focal point of the rift's volcanism (Fairhead, 1976). Lateral separation across the rift, other than that produced by domal uplift, is considered to be small and probably does not exceed 10-20 km in Kenya and Ethiopia.

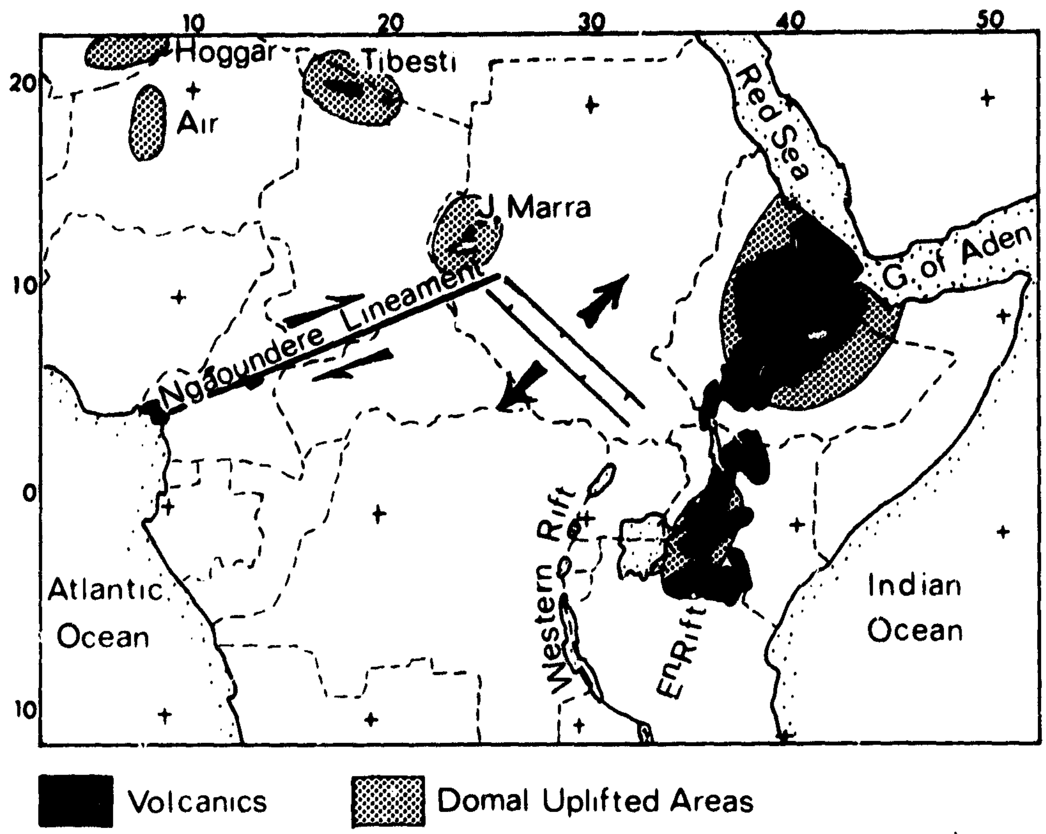

FIGURE 1.

RIFTING WITHIN AFRICA

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Domal uplift and associated volcanism are not restricted to the East African Rift System but occur in the Air, Hoggar, Tibesti and Jebel Marra regions of North Africa (see Fig. 1). Existing gravity data indicate these regions are linked and may represent an earlier stage in the surface expression of continental disruption than the East African Rift System (Fairhead, 1979). To investigate the lithospheric structures associated with these volcanic uplifts a regional geophysical investigation of Jebel Marra was initiated in 1980 (see Bermingham et al., 1981). Compilation of regional gravity for Central Africa suggests that the uplifted Jebel Marra volcanic centre forms a third arm of an intraplate triple junction. The other two arms are the WSW trending Ngaoundere (or Foumban) lineament that can be traced 2000 km from W. Sudan through Central African Republic, Chad into Camerouns and the SE trending rift located entirely within Western and Southern Sudan. These two arms are subsiding sedimentary rift features and it has only been geophysical studies (e.g. Louis, 1970) that have so far revealed their tectonic (and economic) importance. The Ngaoundere lineament is a dextral shear zone of late Pan-African origin and can be traced prior to the opening of the S. Atlantic in Brazil as the Pernambuco lineament (Almeida & Black, 1967; Martin et al., 1981). Reactivation of the African section of this shear zone occurred in the Lower Cretaceous at the time of initial formation of the S. Atlantic. This resulted in the development of a series of deep, narrow, fault bounded sedimentary basins in Chad, Central African Republic and Sudan. Further reactivation appears to have occurred in the Tertiary to Pliocene and has faulted and deformed the Cretaceous sediments within these basins with contemporaneous and/or subsequent development of basement uplift and volcanic activity (65-0 Ma) at its eastern and western ends. The SE trending rift in Western Sudan has a similar sedimentary and reactivation history but has been normally faulting indicating predominantly tensional tectonics as indicated by the arrows in Fig. 1. Unlike the East African Rift System there is a lack of seismic activity associated with this rift system deduced from teleseismic studies and local monitoring since April 1980 by the University of Leeds.

Gravity profiles across various sections of the Central African Rift System will be presented to indicate the nature of the regional and residual gravity fields associated with the rift system. Crustal and upper mantle models to account for these gravity fields will be discussed and compared to existing models for the East African Rift System. Similarities between the Jebel Marra and Afar triple junctions suggest that the Jebel Marra triple junction may be an embryonic Afar triple junction. As such it may help to explain the development of such features and why uplift and volcanism are mainly restricted to one arm of the triple junction.

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THE WEST AFRICAN RIFT SYSTEM AND ITS BEARING ON THE ORIGIN OF CONTINENTAL RIFTING

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Continental rifting is usually accompanied by magmatism which persists for long periods, often more than 100 Ma. Since no part of the lithosphere is likely to remain stationary with respect to the asthenosphere for such long periods, this magmatism cannot be the result of hot material rising from the deeper parts of the mantle but is more likely to be the result of the passive upwelling of asthenosphere into the lithosphere. Thus magmatism and the thermal disturbance responsible for it must be caused by rifting and not be the cause of it. Lithosphere stretching, accompanied by asthenosphere upwelling, followed in some instances by thermal relaxation provides a satisfactory model for the formation of continental rift systems and sedimentary basins (McKenzie, 1978; Sclater and Christie, 1980; Brown and Girdler, 1980). The tectonic development of the West African rift system provides strong evidence in support of this model.

The West African rift system as defined here comprises the Cretaceous Benue trough and the Tertiary to Recent volcanic Cameroon line (Fig. 1). The Benue trough is filled with up to 6000m of marine sediments whose deposition was terminated by a period of mild deformation in the Santonian (c.80 Ma). It is often cited (e.g. Burke and Wilson, 1976) as one of the best examples of a failed arm of an RRR triple junction (the other two arms gave rise to the South Atlantic Ocean). At its northeastern end it splits into two smaller rift structures, the Chad and Yola rifts. Volcanic rocks have been reported from the Benue trough but are not extensively developed.

The Cameroon line is a chain of transitional to strongly alkaline volcanoes extending 1600km from the Atlantic island of Annobon, across the African continental margin towards the centre of the continent (Fig. 1). The earliest magmatism on the line is represented by syenite and granite ring-complexes ranging in age from 65 to 35 Ma (Cantagrel et al., 1978). The oldest extrusive rocks have been dated at 45 Ma on the continental sector (P.I. Okeke, unpublished data) and 31 Ma on the oceanic sector (Dunlop and Fitton, 1979). The most recent activity is represented by the active volcano, Mt. Cameroon though recent cones can be found in virtually all parts of the line. The Cameroon line has been almost continuously active over the past 65 Ma and the volcanism shows no consistent migration with time. Volcanism on the continental sector has been accompanied by regional uplift of about 1km but there is no evidence of rift faulting. The siting of the volcanic centres does not appear to have been influenced by the structure of the basement as the line cuts across tectonic lineaments, basement fractures and oceanic transform faults.

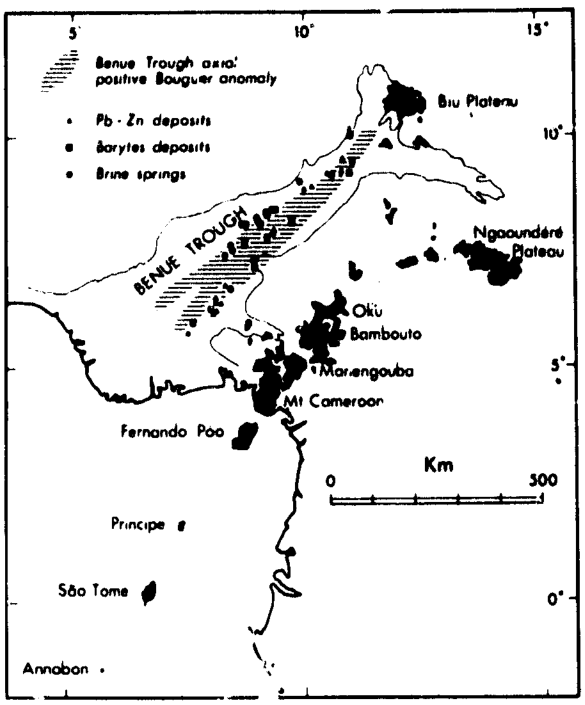

Fig. 1

The West African rift system. Cameroon line volcanic rocks shown black.

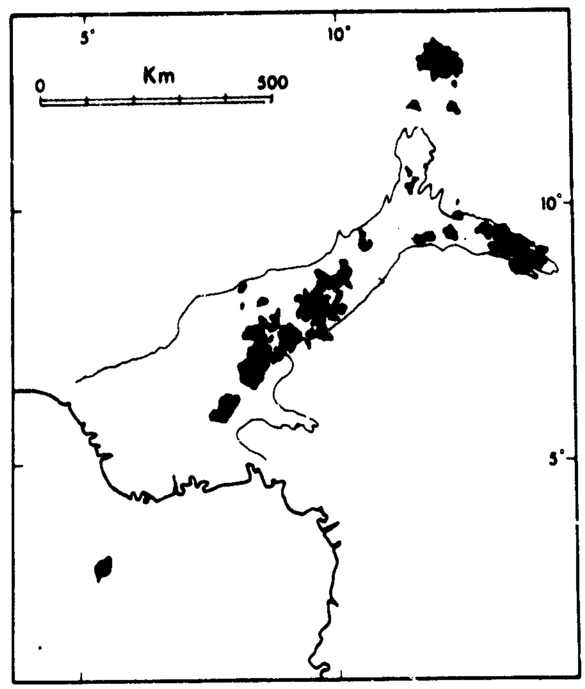

Fig. 2

The Cameroon line superimposed on the Benue trough by rotating the former clockwise relative to the latter by 7° about a pole at 12.2°N, 30.2°E.

(both Figs. from Fitton, 1980)

(Figures published courtesy of Elsevier Sci. Publishing Co.)

An explanation for the origin of the Cameroon line may lie in its relationship with the Benue trough (Fitton, 1980). The two features are so remarkably similar in shape and size that they may be superimposed perfectly by rotating one with respect to the other by 7° about a pole in Sudan (Fig. 2). This geometrical coincidence cannot be accidental but probably results from a displacement of the African lithosphere relative to the underlying asthenosphere. Thus the 'Y'-shaped hot zone in the asthenosphere which would have lain beneath the Benue trough became displaced (relative to the lithosphere) so that it now lies beneath Cameroon and the Gulf of Guinea. The Benue trough and the Cameroon line are, therefore, seen as complementary features. Magmas destined for the Benue rift reached the surface as the Cameroon line instead.

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The displacement required by this model must have occurred between the cessation of activity in the Benue trough (80Ma) and the earliest magmatism in the Cameroon line (65Ma). It is best explained by postulating a short-lived period of rapid clockwise rotation which interrupted the generally anticlockwise rotation implied by the South Atlantic hot-spot traces. Corroborative evidence for such a wobble in the motion of Africa comes from the Walvis Ridge which has been interpreted as a hot-spot trace. This ridge is off-set at magnetic anomaly 34 (79Ma) in the sense predicted by this model and by roughly the required amount. Further evidence that changes were taking place in the motion of Africa at this time comes from the South Atlantic transform faults which change in direction at anomaly 34 (Sibuet and Mascle, 1978).

If this intepretation of the evolution of the West African rift system is correct then it has implications for general models of continental rifting. The West African rift provides a unique example of a rift system interrupted in the course of its development and allows us to examine the effects of the underlying thermal disturbance in isolation from the rifting which produced it.

The Benue trough was produced as part of the much larger rift system which gave rise to the South Atlantic Ocean. This rift system may have been initiated and its course governed by hot spots (Burke and Dewey, 1973) but most probably evolved through stretching of the lithosphere in the areas around and between these hot spots. An inevitable consequence of this stretching would be the development of linear zones where hot asthenosphere welled up passively into the lithosphere. The resulting thermal disturbance would extend down into the asthenosphere. At this point a geological accident decoupled the rift system from the deeper portions of this thermal disturbance and brought it to rest beneath what is now Cameroon. The disturbance could then assert itself in an active role and rise into the overlying lithosphere. This could be accomplished rapidly by convective heat transfer and would have the effect of thinning the lithosphere. In this way an image of the Benue trough thermal disturbance has been imprinted on the lithosphere beneath Cameroon and its magmatic effects are still being felt today.

It is significant that, despite a long history of rift-valley-type magmatism and associated uplift, the Cameroon line has never developed a graben structure. The implication is that graben structures are the surface manifestation of stretching. There is no reason why the Cameroon line should be under tension as the tension which created the Benue trough would have been relieved with the opening of the South Atlantic. The Cameroon line may well be the only example of a truly 'active' (as opposed to 'passive') rift system.

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THE DEEP STRUCTURE AND DYNAMICS OF THE EAST AFRICAN PLATEAU, THE KENYA DOME, AND THE GREGORY RIFT.

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The section of the East African Rift system between 5°N and 10°S is characterised by three main topographic features (Fig. 1): the classical graben structure of the Eastern or Gregory Rift itself; the highland region around the rift valley, known as the Kenya dome; and the elevated area between the Western and Eastern Rifts, known as the East African Plateau (Fig. 2). The topographic similarity between the East African Rift and the mid-ocean ridges was pointed out by Heezen (1959). Subsequent seismic and gravity data led to the notion that the system marks the line of an incipient continental fracture.

From a study of sea-floor spreading rates, transform fault trends, and earthquake slip vectors, Chase (1978) concluded that for the last 5 my the African plate has been moving to the north-east at about 1 cm/yr. Palaeomagnetic data suggests that the plate started its northward movement 100 my ago (Oxburgh and Turcotte 1974) but Briden and Gass (1974) conclude that during the last 40 my, the palaeomagnetic pole did not move systematically relative to Africa. Estimates of extension across the East African Rift system based on global plate motions (Chase 1978) are generally larger than those deduced for geological considerations (Baker and Wohlenberg 1971). Both are small, about 0.6 cm/yr at 5°N, decreasing southwards. The continuity of the seismicity into southern Africa (Fairhead and Henderson 1977) together with the NW-SE tensional stress field derived from focal mechanism studies suggest a direct link with the Gregory and Ethiopian rifts, and by implication the mid-ocean ridge system of the Indian Ocean. This accords with the idea that the East African Rift system is an incipient plate tectonic accreting margin. It could be however that these features are related to a tectonic system involving membrane stresses (Oxburgh and Turcotte 1974). The key may be the seismicity of the region between the Kenyan and Ethiopian Domes where an earthquake recording programme is now in progress.

Rift formation in Africa is related to uplift (Shackleton 1978) and three phases of uplift totalling 1500m since the Cretaceous have been recognised (Le Bas 1971). Volcanism is also associated with the rifting. The understanding of these processes requires a knowledge of the structure of the underlying lithosphere and asthenosphere and some information is available from geophysical studies. Earthquake data suggest that the crustal structure beneath the flanks of the Gregory Rift is of normal continental type (Maguire and Long 1976). Earthquake and seismic refraction data (Griffiths et al.1971) show that the axis of the rift is underlain by anomalous crustal material at shallow depths and an axial intrusion within a few km of the rift floor has been suggested to explain the axial positive gravity anomaly (Searle 1970, Baker and Wohlenberg 1971, Khan and Mansfield 1971). Magnetic variation studies indicate high conductivity material beneath the rift axis (Banks and Ottey 1974) but there is also a deeper more extensive conducting zone beneath Mt Kenya to the east of the rift. Project MAGNET data show a quiet belt coincident with the rift axis (Green 1976), presumably due to the upwelling of the Curie isotherm. Banks and Swain (1978) have used gravity data to reveal anomalies of the compensation of

KENYA RIFT GEOPHYSICS

Khan, M. Aftab and P.K.H. Maguire

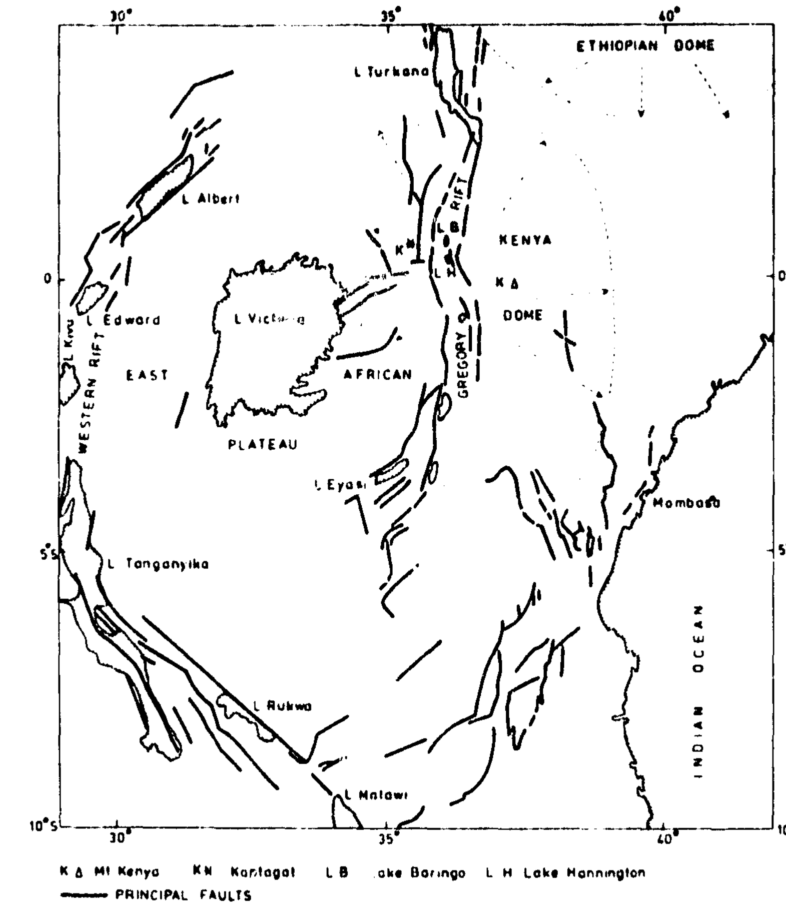

Fig. 1 The East African Plateau, Kenya Dome and Gregory (Eastern) Rift Valley

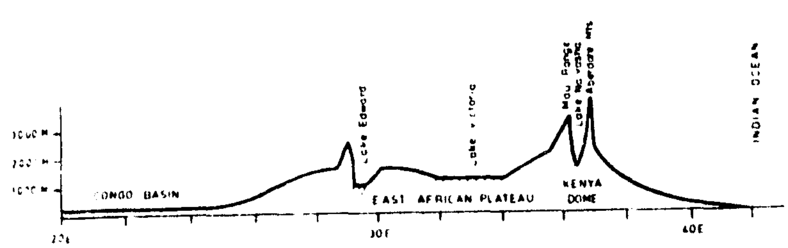

Fig. 2 Simplified topographic profile from the Congo Basin to the Indian Ocean along 0-1°S

(Figures published courtesy of the Geological Society of England.)

the compensation of the topography. Their gravity map shows a north-south trend relating to the plateau and a circular anomaly corresponding to the dome and coinciding with the culmination of the seismic low velocity zone postulated by Long and Backhouse (1976). The interpretation of the available data in terms of lithospheric structure requires control from high resolution and an international programme starting in 1984 is being discussed.

Two principal mechanisms have beenproposed to explain the structure and geology of the East African Rift System. Gass et al. (1978) find that a thermal disturbance at the base of the lithosphere could explain the form and character of the Kenya lithothermal system but the penetrative magmatism cannot satisfactorily explain the volcanics which come from great depths. The second mechanism involves the nucleation of a crack in the lithosphere under tension by lithospheric magma. Both mechanisms may be involved. Once domal uplift and penetrative magmatism have occurred and a tensional stress system sufficient to crack the lithosphere is created, magma from the asthenosphere could make its way to the surface. The magma would come from a depth which depends on the local strength of the asthenosphere, the tensional stresses and the buoyancy forces resulting from the magma-lithosphere density contrast.

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PROCESSES OF PLANETARY RIFTING AS SEEN IN THE RIFTING AND BREAK-UP OF AFRICA

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The East African Rift System was one of the first rift systems to be recognised having been mapped by J. W. Gregory near the end of the last century. It is a very valuable indicator to processes of planetary rifting as nearly all the different forms and stages of rifting have been recognised. These include an evolutionary sequence from insipient rifting in southern Africa progressing northwards to a region of block faulting, major rift valley development with volcanism along the rift floor and culminating in sea floor spreading and the development of a fully oceanic region in the Gulf of Aden. In addition, the rifted Red Sea was enclosed and isolated from the Indian Ocean until the end of the Miocene and provided a complex sedimentary environment with the formation of vast thicknesses of evaporites; its spreading centres provide locations for heavy mineral deposits. Finally, we have the classic riftea shear zone of the Gulf of Aqaba - Dead Sea rift where shear in excess of 100 Km is well documented, this having occurred in two or three stages as the Arabian plate moved northward, rotating anticlockwise, colliding with Asia and forming the Tauros-Zagros mountains

Various geophysical techniques have been used to study the processes of rifting in various parts of the East African Rift System. Seismicity and gravity studies have been particularly valuable in East Africa whilst a combination of magnetic, gravity and seismic techniques have been used in the Red Sea and Gulf of Aden in attempts to estimate the amounts of oceanic crust and their histories.

The rifting in Africa is associated with a long wavelength negative Bouguer gravity anomaly extending for some 5000 Km with a width of 1000 Km. This has been interpreted as being due to attenuation of the lithosphere. The exact way in which this occurs is somewhat enigmatic. In particular, there is no simple relationship of the areal extent of the gravity anomaly to the seismicity, seismically active regions (probably indicative of insipient rifting) occurring where there is no long wavelength gravity anomaly. This would seem to indicate that the rifting occurs first due to tensional stresses in the plate causing brittle fracture. The rifting and extension permits igneous material to rise along "heat leaks" causing the lithosphere to be replaced at its base upwards by the slightly less dense asthenosphere. If this is the case, the rifting is the cause of the attenuation rather than vice-versa.

When we travel north to the Red Sea and Gulf of Aden, the wide negative Bouguer anomaly is less impressive becoming replaced by positive anomalies which begin to evolve along the axis of the rift in Kenya. When corrections are made for the thicknesses of light sediments, the positive Bouguer anomaly is found to

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extend over the whole widths of the Red Sea and Gulf of Aden. The interpretation is the tensional stresses have been large enough to cause complete separation of the old continental lithosphere permitting the evolution of new oceanic lithosphere. Here again, there is evidence that the rifting came first and the uplift of the margins (sometimes referred to as updoming) followed. For example, Azzaroli observes that uplift along the marginal faults followed the deposition of marine sediments, the sediments being disturbed by the faults. Detailed studies of the magnetic and gravity fields over the Red Sea and Gulf of Aden tend to support the view that the continent/ocean boundary is close to the coasts which fit so perfectly. The magnetic anomalies indicate that the sea floor underlying the Gulf of Aden and Red Sea is unlikely to have formed continuously. It seems there has been at least two and very likely three phases of spreading. As the magnetic profiles are short, it is difficult to obtain confident dates for the earlier phases from these alone. Tectonic events from nearby regions such as the Aqaba-Dead Sea rift and the Ethiopian rift have been used to give possible timings and the corresponding synthetic and observed magnetic anomalies compared. In this way, it seems likely that the phases of spreading occurred from 0 to 4.5 My, 16 to 23.5 My and possibly 35.5 to 43 My. Thus the processes of rifting here are discontinuous with hiatuses of the order of 10My.

Clearly, such phenomena have to be taken into account in theories of processes of planetary rifting. The inference that the rifts evolve in a go-stop-go kind of way suggests that the analogy of the movements of the plates with the movements of icebergs floating, getting jammed and freeing themselves might well be appropriate. The stresses set up are mainly horizontal and cause the plates which are thin and fragile (having length to thickness ratios between 50 and 150 to 1) to break by brittle fracture. In this case, the processes we observe in the rifts are second order and there is no simple relationship to the fundamental processes such as convection in the mantle which might be the primary cause of the movements of the plates.

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THE MORTON-BLACK HYPOTHESIS FOR THINNING OF CONTINENTAL CRUST, RE-EXAMINED IN WESTERN AFAR

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The elegant hypothesis of Morton and Black (1975), to explain how continental crust is thinned at young extensional plate boundaries, has found wide appeal in both geological and geophysical circles (eg. Berckhemer et al., 1975; de Charpal et al., 1978). Although the hypothesis was formulated largely on observations made during detailed mapping of southeastern Afar (Black et al., 1972, 1974), it was explicitly applied also to western Afar where estimates of the degree of crustal thinning were made.

A review of published data on the structural geology of western Afar, and specifically the Dessie-Elwà traverse at 12.5°N latitude, shows that faulting and block-tilting of a thick basaltic pile is more complicated than considered by Morton and Black. Just as feeder dikes are concentrated into swarms, so faulting is preferentially concentrated into narrow zones that are not everywhere coincident with the swarms. Thus block-tilting does not increase in magnitude towards Afar in the presumed direction of increasing crustal attenuation, but is steepened at each fault zone (Gortani and Bianchi, 1941, 1973; Mohr, 1971). If there is any overall trend, it is one of <u>dec</u>reasing tilt towards Afar (Gouin and Mohr, 1964). Furthermore, sets of faults and dikes that trend oblique to the main structural orientation of the Afar margin, though subordinate, are not unimportant and indeed hint at the regional stress field operative at the time of margin development. Particularly prominent is an orthogonal set of ENE faulting.

Synthetic faulting at the topographic edge of the plateau is well developed further south towards Addis Ababa, but is absent in the Dessie sector. This is related to the northward dying-out of the Borkenna marginal graben where it intersects a major basaltic shield, the Ardibo massif. The antithetic faults are replaced by a zone of steep tilting and folding. Further east within the margin, the basalt pile exhibits local asymmetric folding and reversed faulting suggestive of compression from the northwest (Gortani and Bianchi, 1973, p. 173; Mohr and Roger, 1966, p. 20).

The evolution of the western Afar margin can be interpreted on the assumption that the original fissure-feeders were vertical and perpendicular to the flood lava surface. Major basaltic volcanism occurred at 25-24 m.y., according to a re-interpretation of the data of Megrue et al. (1972), with an important terminal episode of silicic volcanism (Justin-Visentin and Zanettin, 1974). Block-faulting and tilting occurred immediately following the formation of the volcanic pile, and was certainly completed before the renewed irruption of dikes at 15-11 my ago. No block-tilting was associated with this second dike episode. Subsequent volcanism has been very restricted, but steep normal faulting has occurred through the Quaternary and is still active. This faulting is related to progressive vertical

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Fig. 1 Structural map of the Ethiopian plateau–Afar margin along the Dessie–Mille traverse. Major faults indicated by thick trace, dip of stratoid lavas by arrows. (The Ethiopian plateau extends west from Dessie, the Afar margin lies between Dessie and Eloà, and the Afar floor extends east from Eloà.)

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disjunction between Afar and the Ethiopian plateau, evidenced in the gentle regional tilting of Pliocene-Quaternary sediments, river terraces and erosion surfaces. This latest tectonism can be viewed as the onset of differential subsidence between rift and plateau as the Afar margin crust/lithosphere continues to cool, consistent with westerly dips of older volcanic and sedimentary strata on the floor of Afar adjacent to the western margin.

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MAJOR STAGES OF RIFTING EVOLUTION IN THE EARTH'S HISTORY

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The Earth's history can be subdivided into a number of epochs marked by activization of the uppermost mantle heating (up to its partial or complete melting), horizontal extension of the entire crust or its large parts and origination and rapid development of linear tensional rift-like or rift structures. The concept of the pulsating Earth on the background of its general and irregular expansion seem to explain these epochs' existence most naturally (Milanovsky, 1980). Five major epochs of rifting activization and crustal extention can be distinguished, and among them three epochs with relatively most intensive global extension. They are: Archean, Late Proterozoic and Mesozoic-Cenozoic. Extention zones originated in each of these epochs differ greatly in geometry, morphology, paragenesis of tectonic features, rock formations, thermal regime, also in accompanying magmatism and metamorphism. They also vary in the following geological evolution, in particular, in later deformations occurred in these zones. These differences of the consequent rift-like and rift structure generations testify the irreversible evolution of the Earth and as a part of its changes in the structure and properties of the crust which underwent tension and destruction in various periods of its history.

1. Linear structures of the most ancient Archean epoch of rifting originated on the relatively thin mobile protocontinental crust and combined features of both rift and geosynclinal zones of the following geological time. They were set under conditions of the heat increase, magmatic permeability, destruction and horizontal extension of the most ancient crust; their further development, however, was accompanied by the alternation of regimes from tension to compression (or numerous such alternations) and considerable changes of thermal regime. Probably the earliest of these mobile ancient zones and nevertheless of great "vitality", are zones presented in the recent Earth's structure as charnockite-granulite belts forming a rather large net within ancient platforms of all continents. Further on, these zones underwent numerous tectonic and magmatic regeneration and, in fact, predetermined the localization of late Proterozoic and Phanerozoic rift zones of continents, and the outlines of margins of "secondary" (Mesozoic-Cenozoic) oceanic basins. During periods of more intensive but more local horizontal extension and heating, the Archean protocontinental crust in many places was cut by quite dense net of more narrow and short extention zones presented in the recent tectonic pattern as greenstone belts. These zones filled with thick early and late Archean volcanic and sedimentary materials were recorded as zones of subsequent compression and granitoid

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plutonism. Some scientists consider these zones (greenstone belts) analogue to later eugeosynclinal troughs or island arcs, other-analogue to rift zones. In fact, these belts (like charockite-granulite belts) combined in themselves features of both rift and geosynclinal epochs.

2. Protorift zones of early Proterozoic epoch were set in the result of a partial destruction of a relatively more rigid and mature crust, compared to Archean epoch. They seemed to be related to protogeosynclinal troughs of the same age but differ in scale and role of the following compression deformations (aulacogeosynclinals of the Pechenega-Varzuga-zone type in the northeastern part of the Baltic shield), or sometimes should almost complete absence of any compression manifestations (i.e., Great dyke in Rhodesia). Generation of these protorift structures on the whole is yet poorely revealed and analysed.

3. Sharp intensification of a rifting process is dated back to late Proterozoic (Riphean, especially middle) when numerous continental rift zones-aulacogenes-originated on the ancient, mainly, Laurasian platforms and also on the Australian platform (Amadeus, etc.) (Milanovsky, 1981). By structure and genesis aulacogens related close to the geosynclinal belts of Neogeicums set at the same time, and separated ancient platforms. In the majority of cases aulacogens were branches of these belts penetrating far into the platform bodies or cutting off separate their peripheral segments. In such cases, there originated especially large and deep aulacogens (aulacogeosynclinals) evolution of which was completed by tectonic inversion and folding. In late Proterozoic, a complicated system of aulacogeosynclinal zones originated and evolved within a number of present platforms of the Gondwanaland group (South American, African).

4. Partial regeneration of geosynclinal belts in early and middle Paleozoic was accompanied by regeneration of a number of Riphean aulacogens in the regions of ancient platforms adjacent to these active belts. New aulacogens within the platforms did not seem to originate. By the end of Paleozoic, tension and subsidence in the regenerated aulacogens and in Paleozoic geosynclinal belts ceased and in many of them changed to compression. In some aulacogens it repeated periodically during Mesozoic.

Synchronically with the development of the epiplatform rift zones, in early and middle Paleozoic, in Baikalian and Caledonian folding systems, correspondingly there generated for the first time most early epiorogenic rift zones. They originated in the result of horizontal extension during the Cambrian and, mainly, Devonian time.

First manifestations of the general (Mesozoic) breakup of the Gondwanaland super-continent (super-platform?) can be dated back to late Paleozoic when this process started at its certain

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5. Rifting and accompanied non-geosynclinal magmatic activity reached their gigantic scale during the last Mesozoic-Cenozoic epoch of the Earth's evolution. Rifting of the epoch contrast to Proterozoic and Paleozoic one and with rare exception was not connected directly with the evolution of geosynclinal zones and belts and was not subordinated to it as an associate process within the platforms framing them. On the contrary, it presented the extention and destruction of the crust mainly spatially and genetically connected with the formation of "secondary" oceanic basins. In their peripheral parts along the continental boundaries, perioceanic rift zones and systems were set in Mesozoic, while in their inner parts, intraoceanic rift belts originated since middle Mesozoic. In the latter, horizontal tension and formation of oceanic crust (spreading) reached great scale. Similar belts originated and envolved synchronously within the ancient, though tectonically and magmatically rejuvenated in Mesozoic, Pacific basin. Late Mesozoic rift zones and systems of residual Gondwanaland continents and also Cenozoic rift zones and systems of Africa, North America and Eurasia are of much less scale than intraoceanic and intracontinental rift zones, presenting however, blind branches of these large rift zones, or evolving parallel to them. Contrast to the Pre-Mesozoic rift zones, there is no evidence of the following tectonic inversion and compression in the majority of Mesozoic-Cenozoic continental and oceanic rift zones; in same cases though present they are not distinct and are too local. Mesozoic-Cenozoic rifting is not subordinated to the geosynclinal belts evolution; on the contrary, by its kinematic tendencies it is an "antagonistic" process. These two processes have very complicated relations in space and time and in global scale they balance each other: in time periods of rifting activization alternate with the compression paroxysms in geosynclinal regions during which tension in the rift zones stops or reduces its rate; that leads to the reorganization of their kinematic pattern (Milanovsky, 1980, Schwan, 1980). On the whole, Mesozoic-Cenozoic epoch is marked by predominance of global horizontal extension (in particular, as rifting) over compression regime.

In the Earth's history, thus, rifting as a geological process undergoes complicated evolution. In Archean, development of linear tectonic zones combined features typical of the rifting and geosynclinal processes of the following epochs; during Proterozoic and great part of Paleozoic, the continental rifting was related and subordinated to geosynclinal process, while during the last Mesozoic-Cenozoic epoch, it acquired great independent significance as one of the principal destruction forms, break up and "creeping apart" of continental massives and sea floor spreading in the course of the general activized expansion of the Earth.

Distribution of various types of rift zones and other tectonic regions during the Earth's history is shown on the diagram.

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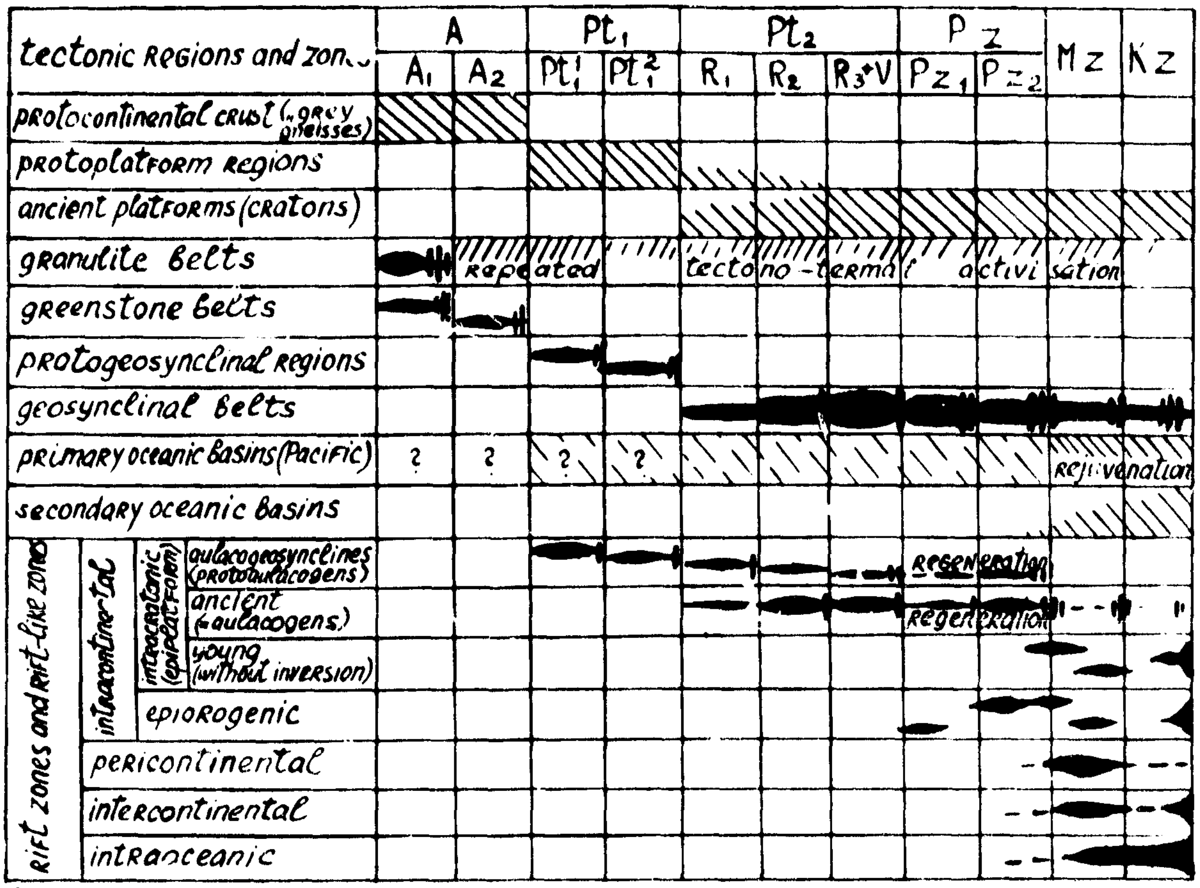

compressive deformation

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INSIGHTS FROM BASIN AND RANGE SURFACE GEOLOGY FOR THE PROCESS OF LARGE-SCALE DIVERGENCE OF THE CONTINENTAL LITHOSPHERE

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Even though the Basin and Range province (BRP) has been cited as a region of Cenozoic extensional tectonics for over a century, only quite recently have we begun to realize our level of ignorance of its structural expresion and magnitude (Davis, 1980). In the last few years, a number of papers (increasing exponentially with time) have identified low-angle normal fault terranes in the BRP, and it is now apparent that their occurence on a province-wide scale is a rule rather than an exception. The belated recognition of so widespread a phenomenon was largely due to the lack of detailed field studies over much of the BRP, combined with the notion held by many early workers that low-angle faulting is mechanically impossible in a tensional regime.

The picture now emerging is that the BRP has been a region of extremely large magnitude extension (hundreds of kilometers) which mostly predates the formation of the modern basins and ranges. The extension is accomodated mostly on large, very low-angle normal faults, termed detachments. Above the detachments are inhomogeneously extended allochthonous masses containing both high- and low-angle normal faults. These allochthons are characterized by thin, nappe-like slices, large- and small-scale rotated fault blocks bounded by both listric and planar normal faults, and large, unextended blocks. Below the detachments lie vast terranes which behaved more or less as rigid blocks while the allochthons were being emplaced.

Consensus among BRP geologists on the relative rigidity of the autochthons beneath the detachments has evolved slowly only over about the last two years. The reason for this is that the allochthons are so highly extended, thin (aspect ratios greater than 20:1), and wide (some exposed over as much as 100 km parallel to their transport directions), that some form of *in situ* accomodation of upper plate extension in the lower plate seemed geometrically and mechanically inescapable to many workers. Further, preliminary data from the southern BRP was very much in support of the *in situ* model, because it was observed that the autochthons contained a regional lineation trending closely parallel to the extension direction in the allochthons. This generated much excitement because for the first time the "classic rift model" could be observed in surface outcrop at all structural levels. In other words, substantiation of most geologist's fundamental concept of continental rifting as large-scale flattening by pure shear (high level "brittle" normal faulting [probably listric] separated from deep-seated "ductile" stretching and dilation by intrusion by a fairly localized brittle-ductile transition zone, Fig. 1) seemed imminent.

However, as the results of detailed studies accumulated, it became clear that most of the southern BRP autochthonous lineations significantly predated the extension and emplacement of the allochthons, and that large areas of autochthons in both the northern and southern BRP did not contain any lineations or other possible source of accomodation. In short, conventional notions of the rifting process and the geometric realities of low-angle extensional faulting in the BRP have very little in common.

The current data base strongly suggests that the detachments are zones of large-scale simple shear highly analogous to thrust faults (Wernicke, 1981), i.e., extension is expressed largely by the divergence of two plates separated

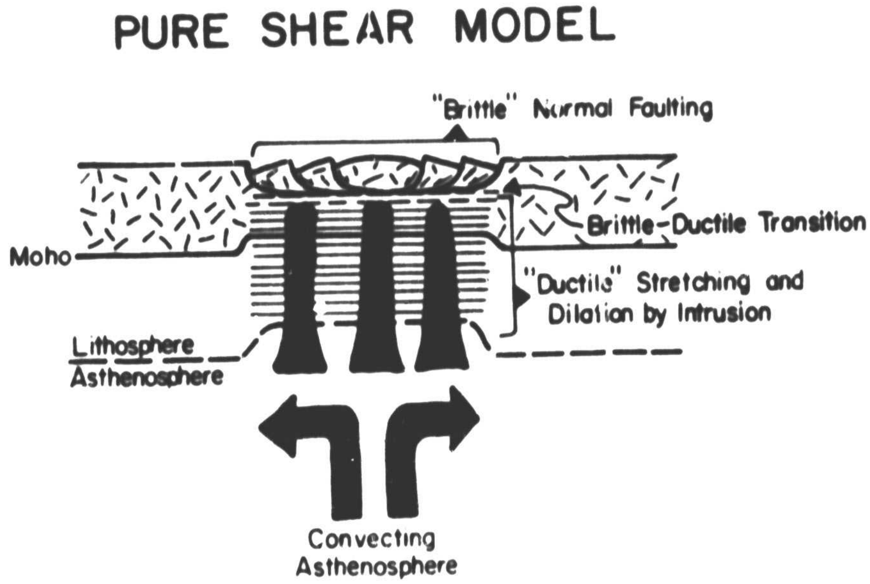

FIGURE 1

by a shallow fault. The fundamental question is, How deep does this style of extension occur in the lithosphere? This question cannot be answered with certainty yet, but for the sake of considering all the possibilities I propose a model in which extension of the lithosphere is dominantly expressed by localized simple shear on a relatively small number of major detachments

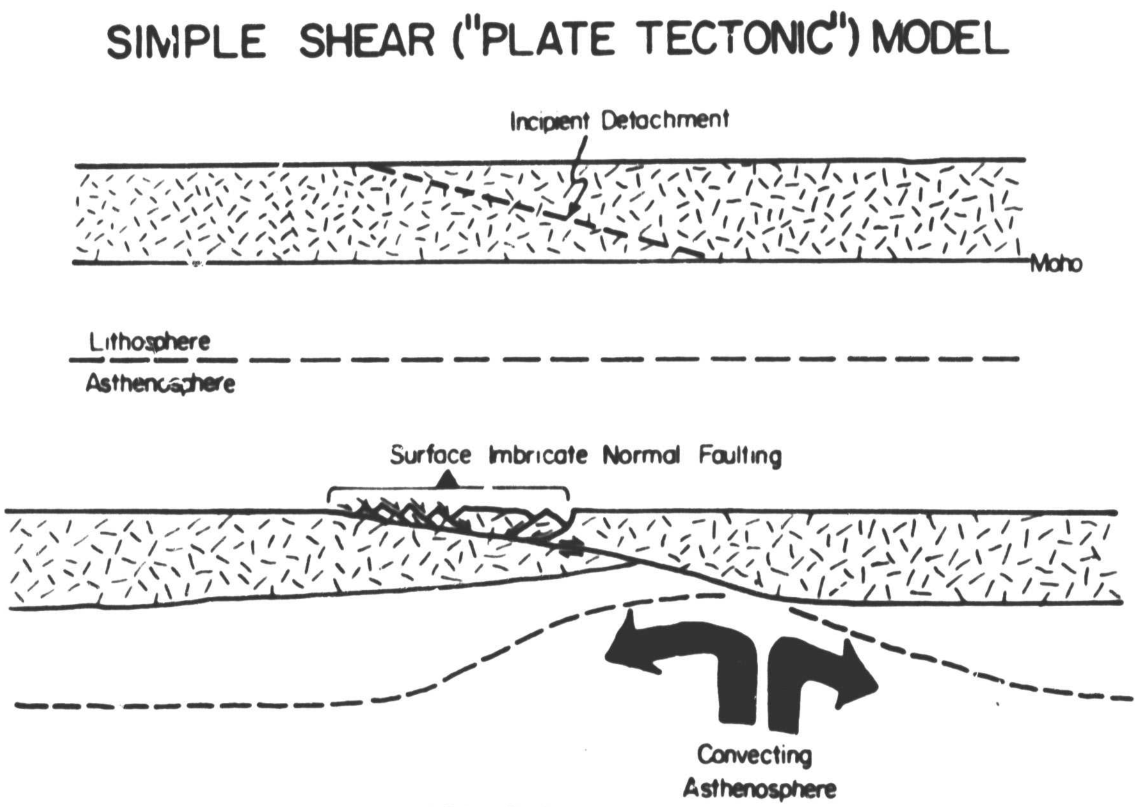

FIGURE 2

(as few as 1) bounding a correspondingly small number of plates (as few as 2, as depicted in Fig. 2). It is entirely possible that elements of both models are correct and vary from rift to rift, but the BRP surface data biases me toward the simple shear end-member.

This "cross section plate tectonics" is potentially very powerful for explaining non-uniform extension between crust and mantle lithosphere, greatly thinned crust which lacks surficial normal faulting, and volcanic centers which are spatially removed from coeval loci of normal faulting. Clearly, the model has major implications for geophysicists who model the rifting process using subsidence histories of post-rift basins.

Postulating rigid lower crust and mantle lithosphere during the rifting process warrants a brief discussion of the terms brittle, ductile, localized, and penetrative. In a rock mechanics laboratory, samples which accomodate strain by the volume increasing process of fracturing mineral lattices are said to be brittle, and those which accomodate strain by the non-volume increasing process of dislocation climb and glide within the lattices are referred to as ductile. At laboratory scales, strain via brittle fracture tends to be localized, and strain via dislocation mechanisms is relatively penetrative. On geologic scales rocks under either brittle or ductile T/P conditions can deform either penetratively or on localized shear planes. In other words, the scale independent, T/P determined deformation modes of mineral lattices are not necessarily an indication of whether or not strain is accomodated in a penetrative or localized manner on a lithospheric scale; the experimentally determined ductile conditions for the lower crust and mantle lithosphere (particularly if heat flow is high) do not argue in favor of taffy-like stretching at those levels.

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THE SUBSURFACE GEOMETRY OF FAULTS AND THE EVOLUTION OF BASINS IN THE NORTHERN BASIN AND RANGE PROVINCE R. Ernest Anderson, U.S. Geological Survey, MS. 966, Denver, CO 80225, and Mary Lou Zoback, U.S. Geological Survey, MS. 77, Menlo Park, CA 94025

Previously published and publicly available seismic reflection data from the northern Basin and Range have been reviewed in an effort to classify the subsurface geometry of major normal faults in the province. The primary criteria established for the classification of the normal faults is the dip of the strata in the sedimentary basin-fill sequence adjacent to the major bounding fault or faults. Specifically, the presence or absence of reverse drag flexing of the basin-fill strata and/or antithetic faulting as well as an associated growth fault pattern of sedimentation was used to distinguish between slip on listric faults and slip on relatively planar, steep normal faults.

Three major modes of faulting responsible for modern basin development have been identified. Faults of each mode are known to be active as evidenced by historic, Holocene, or latest Pleistocene surface rupture. The data indicate that some basins form as relatively simple sags associated with one or more major steep relatively planar normal faults (Figure 1a), as tilted ramps associated with moderately to deeply penetrating listric normal faults (Figure 1b), and as assemblages of complexly deformed subbasins associated with sharply curving shallow listric faults that sole in a detachment surface at relatively shallow crustal levels (4-7 km depth) (Figure 1c). The first type is interpreted to be associated with directly underlying zones of anelastic extension and/or igneous intrusion; the depth beneath the basins to these underlying zones of extension is probably comparable to the width of the broad sags (12-17 km). The second and third types are interpreted to be associated with zones or deep anelastic extension that are laterally displaced from the zones of surface faulting. This lateral displacement depends, in the second mode, on the radius of curvature of the listric faults and, in the third mode, on the dip of the detachment surfaces.

The subsurface data also indicate that as basins mature they grow broader by both structural and sedimentary processes. These processes include: 1) lateral migration of the basin margin away from its axis by outward stepping to successively younger faults, 2) onlap of the downwarped basin-floor ramp by progressively younger strata, 3) coalescence of early-formed subbasins by sedimentary overlap of transverse structural or paleotopographic ridges or longitudinal horst blocks. Although the second and third processes are in a strict sense sedimentary processes of basin growth (in contrast to the first process which is structural), these latter two processes may reflect structural origins. The sedimentary onlapping of the downwarped basin floor ramp probably results from continual sagging of the basin in response to deep anelastic extension. The coalescence of earlier formed subbasins by sedimentary overlap suggests relative quiescence of the faulting responsible for delineation of these subbasins, a possibility which is supported by the general lack of Quaternary faulting internal to the modern basins. Thus, the third process may also be structural-related and tied, as in the first process, to the spacing of currently active faults. The seismic data suggest that these processes of basin enlargement can be associated with any one of the three major modes of basin development outlined above.

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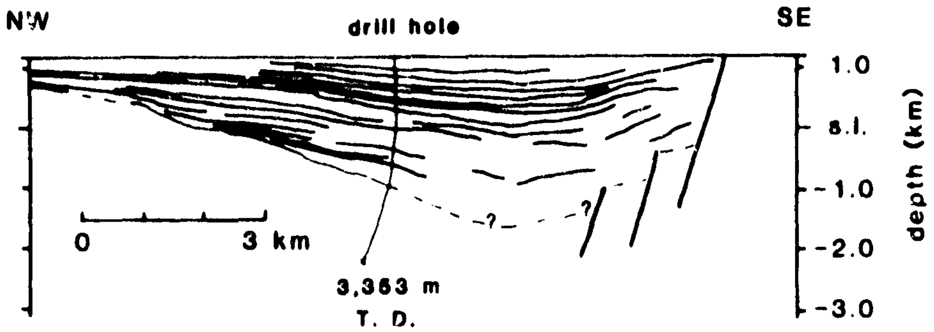

Figure 1a. Line drawing of a migrated depth section across the northern part of the Fallon Basin, Nevada showing major sag-like basin structure (modified from Hastings, 1979). Dashed lines and step-faulting pattern inferred from gravity data.

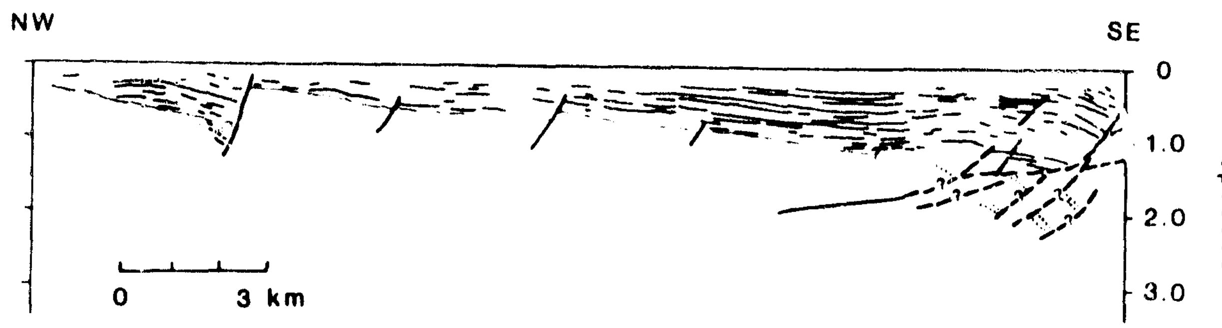

Figure 1b. Line drawing of a seismic section across Mary's River Valley, Nevada showing listric fault-tilted ramp structure (modified from Effimoff and Pinezich, 1981).

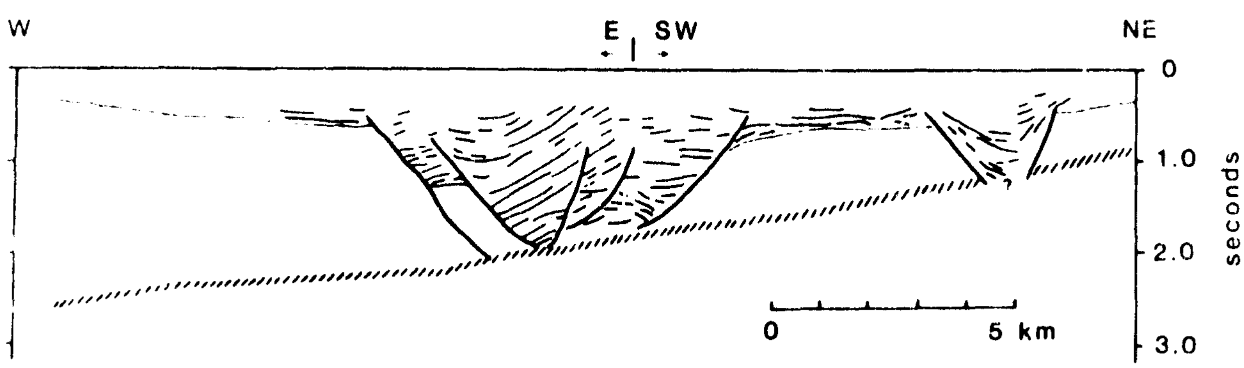

Figure 1c. Line drawing of a seismic section across the Sevier Desert basin, Utah showing sharply listric faults above a relatively shallow detachment surface (modified from McDonald, 1976).

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A corollary to the aspect of basin enlargement seems to be that as basins grow broader or coalesce the deformation style becomes less complex. Early subbasins may display opposed stratal tilt directions either along strike or across separating intra-basin horst blocks. Also, the subbasins may represent contrasting times and rates of formation. The structural complexity of early-formed, and presently deeply buried parts of basins is generally not evident in the surface and near-surface stratigraphic and structural patterns. The relative structural simplicity of the broad mature basins suggests a widening and simplifying of strain domains with time. These evolutionary trends reflect not only a broadening of basins but a broadening of the spacing between the faults that play an active role in basin development. The trends may be part of a pattern of cooling and strengthening of the crust with time.

Development of modern Basin and Range physiography with a characteristic width of major range and basin blocks of 30 km is a relatively recent event in the extensional deformation of the Basin and Range province. Where dated, the formation of the modern basins and ranges in the northern Basin and Range (true basin-range extension) is found to be generally younger than 10 myBP and in places younger than 7 myBP (Zoback, Anderson, and Thompson, 1981). In contrast, accumulating data suggest extensional deformation (pre-basin-range extension) was underway locally by at least 36-30 myBP (Gans, 1981; Zoback, Anderson, and Thompson, 1981). Evidence for this early extension is now found in faulted and highly tilted strata exposed in uplifted range blocks, by large regions of the crust underlain by passively emplaced subvolcanic batholiths, and by the thickness and distribution of stratigraphic units. The possibility exists that in some areas extension on widely spaced, steep, deeply penetrating faults (basin-range faulting) is a late-stage process of extensional deformation that evolved from earlier (Miocene in many areas) large-magnitude thin-skinned extension (characterized by shallow listric and detachment faults) that developed near sites of strong thermal (magmatic) disturbances (including the so called "metamorphic core complexes").

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THE WHIPPLE-BUCKSKIN-RAWHIDE DETACHMENT TERRANE, CALIFORNIA AND ARIZONA: WHAT DOES IT TELL US ABOUT LARGE-SCALE SUPRACRUSTAL EXTENSION?

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The recent recognition in the Basin and Range Province, western U. S. of large areas (>10,000 km^2) of extreme extension (>100%) associated with low-angle normal faults has led to much speculation on the nature and geometry of deformation in these highly extended terranes. The most favored modes of deformation in these terranes have been gravity sliding (Hunt and Mabey, 1966; Moores and others, 1968; Shackelford, 1980) or brittle extension of the upper crust concurrent with ductile stretching and/or igneous intrusion of the lower crust (Anderson, 1971; Davis, 1980; Rehrig and Reynolds, 1980). Most recently Wernicke (1981) has suggested that the master low-angle normal fault represents a rooted normal fault with upper plate extension being the result of shear between large coherent sheets.

The geometry of faulting in these terranes is of critical importance to our understanding of the extensional process. Are the normal faults planar or listric? Moreover, during extension did the rocks behave like tilted dominoes (domino faulting of Chamberlin, 1970) or were more complex processes involved?

The Whipple-Buckskin-Rawhide detachment terrane, southeastern California and west-central Arizona is an excellent area to study some of the above problems. The Whipple-Buckskin-Rawhide terrane is part of a regional (>15,000 km^2) low-angle normal fault complex in southeastern California and western Arizona (Davis and others, 1980, Rehrig and Reynolds, 1980; Frost and others, 1981). This terrane is everywhere underlain by a subhorizontal low-angle normal fault (detachment fault) that separates a lower plate assemblage of Precambrian to Mesozoic or lower Cenozoic igneous and metamorphic rocks and their mylonitic equivalents from an allochthon of Precambrian to Tertiary rocks.

In the Whipple-Buckskin-Rawhide terrane rock units consistently dip to the southwest, rotated along NW-trending, NE-dipping normal faults that sole into the Whipple-Rawhide fault, the master detachment fault in this terrane. Allochthonous units were displaced to the northeast with displacements in excess of tens of kilometers indicated in the Buckskin and Rawhide Mtns.

Upper plate geometry, hence structure, is less complex in the Whipple Mtns. (more proximal portion of terrane) than in the Rawhide Mtns. (more distal portion of terrane). In the Whipple's the upper plate essentially consists of a discrete set of tilted fault blocks. Some of the normal faults that bound these blocks are decidedly listric in geometry, however many of the faults appear to be quite planar, flattening only within a few feet or tens of feet from the Whipple fault (Frost, 1981). The faults may branch and interconnect, yet many fault blocks maintain a fairly uniform and individual character, hence resembling a tilted deck of cards or dominoes.

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The upper plate in Rawhide Mtns. consists of a seemingly chaotic assemblage of rocks dismembered by a myriad of high- and low-angle normal faults. There appear to be two major upper plate structural sheets, a lower sheet consisting of Paleozoic metasedimentary, Mesozoic(?) igneous and metavolcanic, and Tertiary sedimentary and volcanic rocks and an upper sheet of Precambrian(?) crystalline rocks unconformably overlain by Tertiary sedimentary and volcanic rocks. Structure in the upper sheet is relatively simple, consisting primarily of two large tilted fault blocks underlain by a low-angle normal fault whereas structure in the lower sheet is exceedingly complex; all rock units are separated by low-angle faults and are highly deformed internally. Both listric and planar normal faults are present.

Extensional tectonism began in the Whipple-Buckskin-Rawhide terrane in the Oligocene (≈ 30 m. y. BP) and continued intermittently (episodically?) until the mid-Miocene (≈ 10 m. y. BP). Syntectonic deposition of Tertiary units is indicated by the presence of numerous monolithic breccias and conglomerates, while fan-like dips of Tertiary strata record synchronous faulting and sedimentation; older Tertiary units are rotated more than the younger Tertiary units. In addition, older faults are rotated more (have shallower dips) than younger faults.

The presence of clasts of mylonitic gneiss in the Tertiary sediments, occurrence of mylonitic rocks in the allochthon, pronounced discordance of the Whipple-Rawhide fault with foliation in the lower plate, the apparent absence of mid-Cenozoic plutons, and the widespread occurrence of non-mylonitic Precambrian and Mesozoic(?) igneous and metamorphic rocks in the lower plate argue against formation of this terrane by brittle upper crustal extension synchronous with ductile deformation and/or igneous intrusion of the lower crust. In addition, a gravity slide origin for this terrane is made less tenable by the vast regionality of the complex and the apparent absence of the required zone of shortening (toe) to accommodate the large-scale extension.

The Whipple-Rawhide fault was warped into a series of NE-trending antiforms and synforms which presently define the anomalous NE-trend of some of the ranges in this area. Regionally the fault dips to the northeast. When last seen in the Rawhide Mtns. the fault dips about 5°-10° to the northeast and projects under the adjacent Artillery Mtns. and Date Creek Basin. Geologic and geophysical data suggest that the fault can be projected to the northeast under the southwest edge of the Colorado Plateau. If this geometry is correct it requires that the Colorado Plateau be part of the upper plate of a regional low-angle normal fault complex that is rooted in the crust beneath the Plateau, perhaps in a manner similar to that proposed by Wernicke (1981).

Our detailed knowledge of the geometry and kinematics of deformation in the Whipple-Buckskin-Rawhide terrane provides important data toward our understanding of the extensional process. The regionality of this and associated low-angle normal fault terranes requires (1) the recognition of the importance of low-angle normal faulting in Basin and Range tectonics and (2) a radical rethinking of our ideas about the timing and nature of extensional tectonics in the western U. S.

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RIO GRANDE RIFT: ACTIVE OR PASSIVE? EVIDENCE FROM DETAILED EVENTS IN THE TECTONIC HISTORY OF THE ESPAÑOLA BASIN, AND RELATION TO THE STATE OF STRESS IN THE BASIN AND RANGE PROVINCE OF THE WESTERN UNITED STATES

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The Rio Grande rift is an active intraplate rift that trends NNE from south-central New Mexico to Colorado. The borders of several major physiographic provinces lie along the Rio Grande Rift: the Great Plains and Southern Rocky Mountains to the east, and the Colorado Plateau and Basin and Range to the west. Understanding the relationships between these physiographic provinces and the Rio Grande rift is of major importance in unravelling the tectonic evolution of the western United States. Also of major importance in understanding the processes of rifting is determining whether or not a rift can be classified as "active" or "passive" according to the definitions of Sengör and Burke (1978) and Baker and Morgan (1981). The active model suggests that rifts are a result of convection in the asthenosphere or mantle which creates uplifts that, in turn, cause rifting to occur. The passive model suggests that the mantle is passive and that rifts result from faulting in the lithosphere that leads to the development of a local mantle uplift. Thus the timing of events in the evolution of a rift should be diagnostic of the process responsible for rifting. This paper will present evidence from the Española basin of the Rio Grande rift that suggests a direct relationship between a change in the state of stress in the Basin and Range province of the western United States and the onset of significant events in the evolution of the Española basin. This relationship, in part, supports a passive type rifting process for the Rio Grande rift.

The Española basin is the middle of three N-trending, right-echelon basins that make up the central Rio Grande rift in northern New Mexico. The trend, position, and geometry of these basins appear to be controlled by the older bordering Laramide uplifts of the Southern Rocky Mountains. The Española basin is a fault bounded valley about 60 km wide with a thick sedimentary fill that generally dips westward (Fig. 1). The type area of the basin-filling, Miocene Santa Fe Group is exposed in the central and eastern parts of the Española basin. In the western part of the Española basin these sediments are overlain by the Jemez volcanics with radiometric ages from 10 to 0.4 m.y. ago (Dalrymple et al., 1967; Doell et al., 1968).

The presently active subbasin within the Española basin is the central Velarde graben (Fig. 1), which began forming in early Pliocene time (Manley, 1979). The western edge of the Velarde graben is formed by the Pajarito fault zone which cuts the eastern flank of the Jemez Mountains and vertically displaces the 1.1-m.y.-old Tshirege Member of the Bandelier Tuff by as much as 100 m (Golombek, 1981a). Mapping of the Pajarito fault zone (Golombek, 1981b) has revealed the detailed events in the latter half of the evolution of the central part of the Española basin. Evidence upon which the following geologic and tectonic history is based can be found in Golombek (1981b).

The central Rio Grande rift began forming ~27 m.y. ago; however, the early rift was probably similar to a broad downwarp rather than a steep sided

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graben (Baltz, 1979). In the central Española basin, syn-rift Santa Fe Group sediments were deposited directly on flat-lying, pre-rift, Eocene sedimentary rocks without intervening structural disturbance. Localized faulting, concomitant with sedimentation, created central subbasins, indicated by Bouguer gravity minima (Budding, 1979), that contain 2-2.5 km of low density sedimentary fill (Santa Fe Group). These faults became inactive before 10 m.y. ago when Santa Fe Group deposition ended in the central Española basin.

Jemez Mountain volcanism began by ~10 m.y. ago. West-tilting of old volcanics and basin fill occurred from 8.5-7.5 m.y. ago due to movement on faults that define the western border of the Española basin. Volcanism continued after tilting, forming large volcanic constructs which shed large volumes of volcaniclastic sediments. Epeirogenic uplift elevated the entire northern Rio Grande rift over 1 km from 7 to 4 m.y. ago (Axelrod and Bailey, 1976). The Pajarito fault zone and central Velarde graben formed ~5 m.y. ago with subsequent vertical displacement of 200-600 m. The last volcanic activity was 0.4 m.y. ago with subsidence continuing to the present.

Recently Zoback et al. (1981) were able to document a change in the orientation of the least principal stress (extension direction) in the Basin and Range province of the western United States from WSW-ENE prior to 10 m.y. ago to WNW-ESE to E-W from 10 m.y. ago to the present. The earlier WSW-ENE orientation formed the southern Basin and Range province between 20 and 10 m.y. ago; the later WNW-ESE to E-W orientation resulted in the N- to NE-trending structural grain of the northern Basin and Range. This rotation in minimum stress direction ~10 m.y. ago is believed due to right-lateral shear imposed on the western United States by the development of the San Andreas transform fault. The change in extension direction could have 2 overall effects on the NNE-trending central Rio Grande rift. 1) Greater strain rate or rate of extension across the central rift and 2) increased movement on existing NNE-trending normal faults and preferential development of new NNE-trending normal faults.

Consequences of these effects can be seen in the tectonic history of the Española basin. Prior to 10 m.y. ago, Sante Fe Group was deposited in broad downwarped basins (Baltz, 1979). Although some faulting probably occurred, most was localized in the center of the basin. Thus, the initial development of the Española basin did not involve any pervasive faulting, tilting, or folding. The present western margin of the Española basin formed as a result of faulting that began ~10 m.y. ago (Manley, 1979), and these N- to NE-trending faults project directly beneath the center of the Jemez volcanic field (Fig. 1), which also became active at that time (~10 m.y. ago). Movement on the western border faults also caused the west-tilting of old volcanics and sedimentary fill within the Española basin 7.5-8.5 m.y. ago. The reorientation of the regional extension direction from WSW-ENE to WNW-ESE would favor development and movement of these faults. Furthermore, the entire northern Rio Grande rift underwent epeirogenic uplift of over 1 km that began about 7 m.y. ago (Axelrod and Bailey, 1976). Finally, the NNE-trend of the Pajarito fault zone and Velarde graben which formed ~5 m.y. ago is consistent with the proposed WNW-ESE extension direction (Fig. 1).

The above discussion shows that the general sequence of events in the tectonic evolution of the Española basin is: main faulting (rifting) around 10 m.y. ago followed by uplift and doming of the northern Rio Grande rift 7-4 m.y. ago. This sequence of rifting preceeding uplift suggests a passive rifting process. Other characteristics also support a passive rifting process for the Rio Grande rift. It has long been recognized that the character,

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trend, and geometry of the northern Rio Grande rift appear to be controlled by the older bordering Laramide uplifts of the Southern Rocky Mountains, as would be expected if the rift were a result of lithospheric extension rather than asthenospheric convection (Baker and Morgan, 1981). Furthermore, the fact that rifting in the Española basin is related to the regional state of stress also suggests that extension in the lithosphere is controlling and modulating the process of rifting.

Thus, specific events in the detailed tectonic history of the Española basin appear to be directly related to the change in state of stress in the Basin and Range province thereby linking the development of the Rio Grande rift with the rest of the western United States. This coupled with the finding that the main stage of rifting preceeded uplift suggests that the Rio Grande rift is the result of a passive rifting process.

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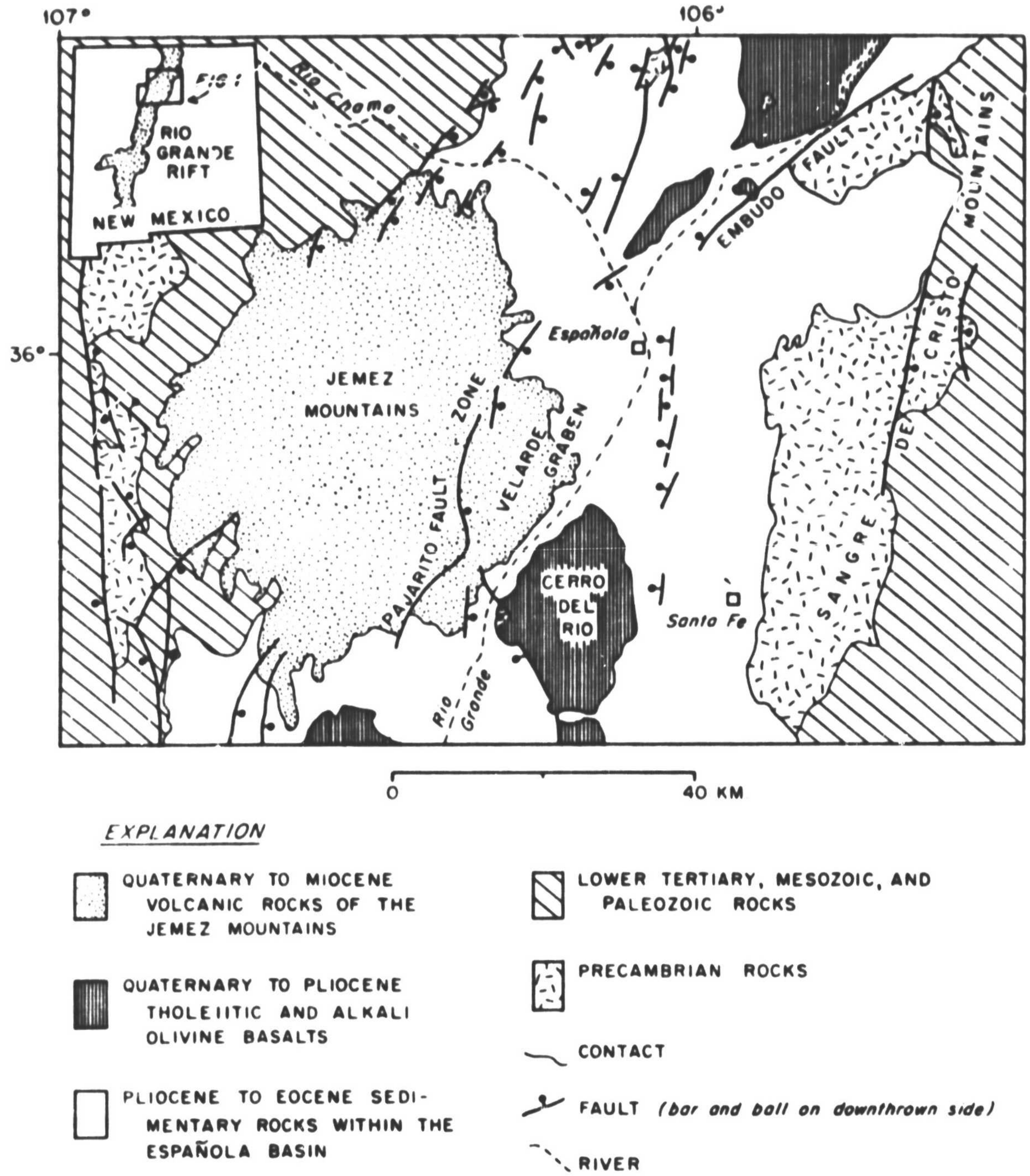

Figure 1. Generalized geologic map of the Española basin of the Rio Grande rift, New Mexico (from Manley, 1979, with modifications). The eastern edge of the Española basin is marked by both the depositional and faulted contact of the basin fill (Santa Fe Group) with the Precambrian rocks of the Sangre de Cristo uplift. The western margin is marked by a series of NE-trending, down-to-the-east faults that are partially covered by young volcanics of the Jemez Mountains.

CRUSTAL STRUCTURE OF RIFTED CONTINENTAL MARGINS: GEOLOGICAL CONSTRAINTS FROM THE EARLY PROTEROZOIC (1.9 Ga) AKAITCHO GROUP, WOPMAY OROGEN, NORTHWEST TERRITORIES, CANADA. R.M. Easton, Department of Geology, Memorial University, St. John's, Newfoundland, CANADA A1B 3X5.

Present understanding of the early development and crustal structure of rifted continental margins is poor because rift structures of modern continental margins are commonly covered by thick sedimentary sequences. This difficulty can be overcome by studying sediment-starved continental margins, very young rifts, or the products of rifting preserved in deformed continental margins. In deformed margins, the ability to observe strata in cross-section compensates for the deformation and metamorphism that the rocks have undergone. Such a deformed rifted margin is located in the early Proterozoic (1.9 Ga) Wopmay Orogen (Hoffman, 1980) in the Bear Structural Province in the northwest Canadian Shield. This paper describes the products of rifting of this margin, and its geological evolution. Comparison of the geological history of the Proterozoic rifted margin with recent margins lends support to the view of Talwani et al. (1979) that the crustal structure of rifted continental margins is unique.

Wopmay Orogen records the development of an early Proterozoic continental margin and its destruction, i.e. a complete Wilson Cycle (Hoffman, 1980). Two rock groups record the initial development of the Wopmay continental margin. The oldest, the Akaitcho Group is a sequence of bimodal (subalkaline basalt and rhyolite) volcanic rocks and arkoses deposited in a rift. The Akaitcho Group is conformably overlain by a passive continental margin sedimentary sequence (Epworth Group) consisting of a lower quartzite sequence and an upper carbonate bank. To the east, the Epworth Group rests unconformably on Archean basement of the Slave craton. A reconstruction of the Wopmay continental margin prior to its destruction is shown in Figure 1.

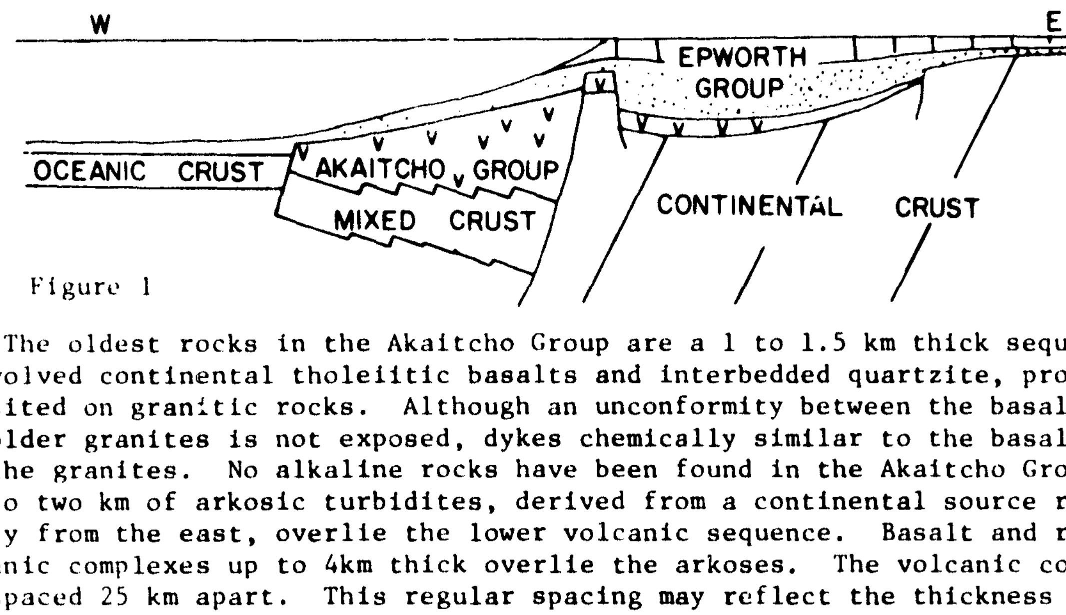

Figure 1

The oldest rocks in the Akaitcho Group are a 1 to 1.5 km thick sequence of evolved continental tholeiitic basalts and interbedded quartzite, probably deposited on granitic rocks. Although an unconformity between the basalts and the older granites is not exposed, dykes chemically similar to the basalts do cut the granites. No alkaline rocks have been found in the Akaitcho Group. One to two km of arkosic turbidites, derived from a continental source region, mainly from the east, overlie the lower volcanic sequence. Basalt and rhyolite volcanic complexes up to 4km thick overlie the arkoses. The volcanic complexes are spaced 25 km apart. This regular spacing may reflect the thickness of the crust in the rift at this time. The volcanic complexes consist of 1 to 3 km of pillow basalt, capped by up to 1 km of rhyolite, mainly extruded as subaqueous lava domes and tuff, but locally subaerially deposited rhyolite is present. Rhyolites are subalkaline, and range from 69-76% SiO_2. Basalts range from 48 to 53% SiO_2, and no intermediate rocks are present. Felsic/mafic ratios in the volcanic complexes are about 20 to 30. Thick rhyolite sills coeval with the extrusive rhyolites intruded the underlying arkosic turbidites.

REE patterns (Figure 2), high Sr^{87}/Sr^{86} ratios, granitoid xenoliths and large volumes of rhyolite indicate that the rhyolites were crust-derived. The major, trace and REE geochemistry of the basalt portion of the volcanic complexes show an evolution upsection from continental tholeiites to ocean tholeiites (Figure 2). In addition, the more westerly basaltic rocks (i.e. further from the Wopmay shelf edge) evolved more rapidly, and completely to ocean tholeiites. The geochemical evolution of the Akaitcho Group basalts from continental tholeiites to ocean tholeiites can also be seen throughout the complete Akaitcho Group section (Figure 2), not only the individual volcanic complexes. To the east, basalt and minor rhyolite lie on the craton, and underlie the quartzite sequence of the Epworth Group (Figure 1). This indicates that the shelf edge was not a major barrier at this time, and that significant subsidence had occurred in the basin relative to the craton in order to accommodate the volume of rift volcanics and sediments. The volcanic complexes are overlain by 1 to 3 km of pelite derived mainly from erosion of the volcanic complexes, but also from the craton to the east. A later period of tholeiitic basalt magmatism resulted in the injection of gabbro sills into the pelites and local accumulations of pillow basalt. The stratigraphic section above the volcanic complexes is similar to sections drilled in the Gulf of California on DSDP Leg 64 (Curray et al. 1980) (Figure 3). Final foundering of the rift then occurred, and the Akaitcho Group was buried by deep water pelites of the Epworth Group. The deep water pelites are continental rise deposits west of the known shelf edge of the Wopmay continental margin. The Akaitcho Group is not unique, Wardle (1981) has described a similar rift sequence in the 1.8 Ga Labrador Trough in the eastern Canadian Shield.

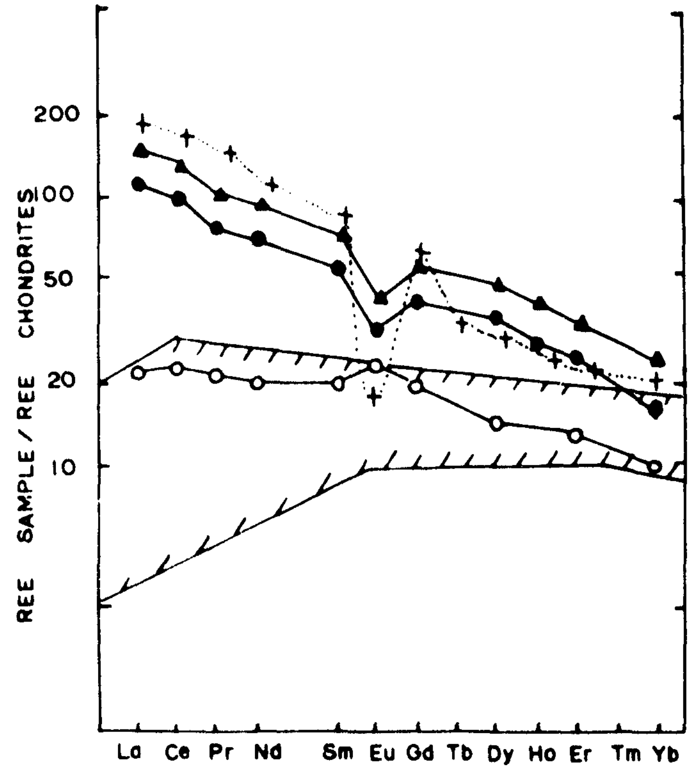

Figure 2: REE patterns of Akaitcho Group volcanic rocks. Crosses - rhyolite. Triangles are from the lower basalt sequence, filled circles from the top of a basalt complex. Note that within the Group, and a single complex, a change from continental tholeiites to ocean tholeiites with younging is observed. About 1 to 2 km of rock lies between each of the samples shown. Field is range of MORB basalts from the mid-Atlantic ridge at 22°S.

CRUSTAL STRUCTURE ... AKAITCHO GROUP

Easton, R.M.

The presence of granitic rocks below the lower volcanic sequence, and the abundance of crust-derived rhyolites in the middle Akaitcho Group points to the continental character of the crust underlying the Akaitcho Group. The spacing of the volcanic complexes may also indicate that the crust was relatively thick (25 km). On the other hand, the evolution of the basalt chemistry from continental to ocean tholeiite both with younging and distance from the shelf edge indicates that some 'oceanic' rocks were present. Thinning of the crust was also necessary if only to accommodate the volume of Akaitcho Group strata (minimum 6 km), and to allow the upper Akaitcho Group to spill over part of the craton. Injection of basalt into the crust in sufficient quantities to cause generation of the large volumes of crust-derived rhyolite, in conjunction with thinning of the crust, must have caused substantial changes to the original continental character of the crust. The subsequent history of the Wopmay continental margin and its destruction (Hoffman, 1980) also argues for the formation of ocean crust to the west of the Wopmay continental margin. The Akaitcho Group thus preserves part of the crust which lay between the continent to the east, and ocean to the west (Figure 1), but the crust beneath the Akaitcho Group cannot be regarded as either oceanic or continental.

The stratigraphy of the upper Akaitcho Group is similar to that found in present day rifts such as the Gulf of California, and the stratoid series in Afar, and to aborted rift sequences such as the late Proterozoic-early Paleozoic Burin Group in eastern Newfoundland. The crustal structure of these modern rifts (Figure 4) is abnormal, being neither truly oceanic or continental in character. The apparent mixed character of the crust under the Akaitcho Group, and its stratigraphic similarity to modern rifts lends support to the view of Talwani et al. (1979) that the crustal structure of rifted continental margins may be unique.

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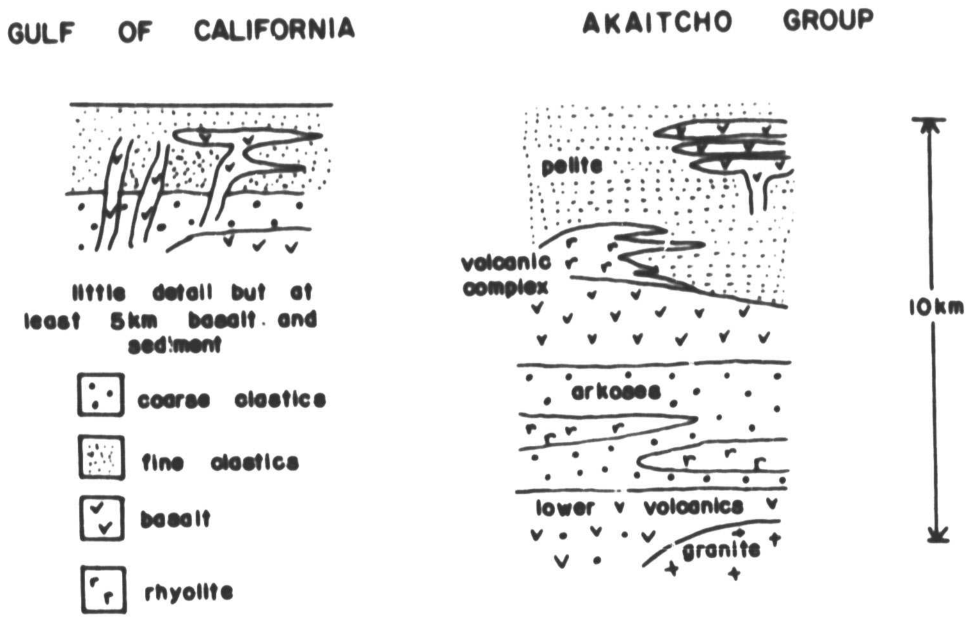
Figure 3: Comparison of the stratigraphy of the Akaitcho Group with the Gulf of California. Data from Einsele et al. (1980), Curray et al. (1980).



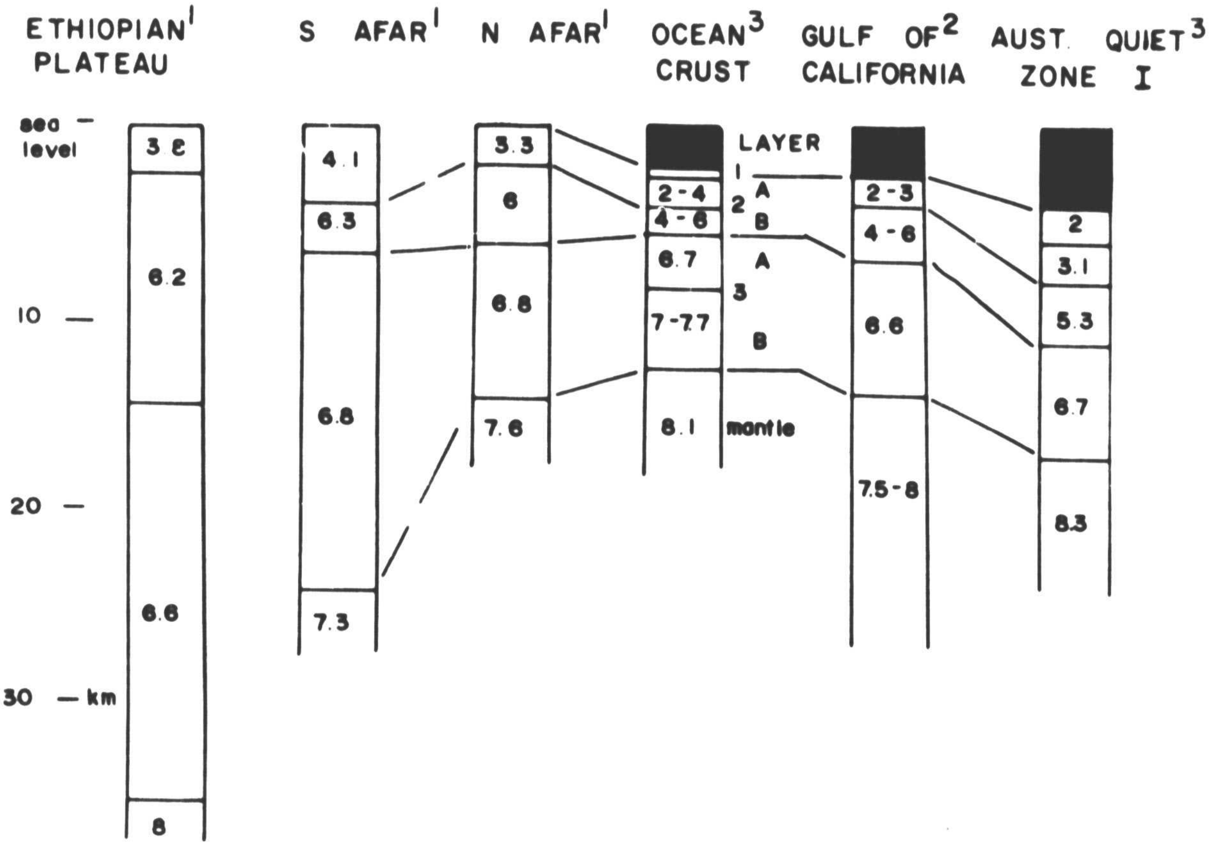


Figure 4: Crustal sections for several modern rift areas. Data from 1) Mohr, 1978; 2) Moore, 1973; 3) Talwani et al. 1979.

A PRELIMINARY ACCOUNT OF THE NEWARK RIFT SYSTEM

by

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The Newark rift system along the eastern edge of the North American continent is thought to represent an early stage in the events that separated the North American plate from the Eurasian and African plates. Processes associated with rifting are considered to be responsible for potentially lucrative accumulations of coal, petroleum, gas, oil shale, phosphate, uranium, zinc, and copper deposits.

The Newark rift system (fig. 1) encompasses the faults, grabens and tilted fault-block basins filled with terrestrial deposits, and extrusive tholeiitic volcanic rocks of Late Triassic and Early Jurassic age. Six of the exposed basins are longer than 100 km. Deep drilling, and seismic, gravity, and aeromagnetic surveys have revealed the location of similar basins onshore and nearshore. The relationship between the onshore basins and more than 60 basins under the Atlantic Continental Shelf is obscured by the lack of sufficient drilling.

The Newark rift system consists of a sequence of individual rift valleys that extend from Nova Scotia to Georgia, a distance of at least 2,300 km. It may extend as far north as the southern North Sea of western Europe and south to the Gulf of Mexico. Individual rift valleys contained ancient lakes, some of which were as large (Newark-Gettysburg-Culpeper basin) and possibly as deep (Dan River-Danville basin) as present-day Lake Tanganyika. When lakes were present, they existed in grabens as does Lake Albert, or along tilted fault blocks as does most of Lake Tanganyika. The remnants of the largest exposed basins have been named the Fundy, Hartford, Newark-Gettysburg-Culpeper, Scottsville, Taylorsville, Richmond, Farmville, Dan River-Danville, and Deep River (Durham-SanfordWadesboro) basins. As in all rift systems studied on Earth, the Newark rift system was activated along pre-existing zones of weakness which can be recognized by the presence of shear zones in Proterozoic or Paleozoic rocks.

Today, strata in the exposed and concealed basins are tilted and the dipping beds have been truncated. Post-depositional tilting coupled with later erosion have eliminated one entire margin of most of the basins, making it difficult to determine whether the remaining segments were once grabens or tilted fault blocks. The thick lines on figure 1 show where faults have been recognized, usually by the presence of talus-slope fanglomerates on the margins of the basins.

Pollen, spores, and the remains of numerous animals date the filling stages from middle Carnian to Pliensbachian(?). Palynomorph-bearing beds, associated with extrusive tholeiitic basaltic lavas in the northern half of the Newark rift system, date volcanism from late Rhaetian (latest Triassic) through Early Jurassic. Evidence for earlier volcanism in the Late Triassic is elusive. For example, masses registering as both gravity and magnetic highs suggest large volcanoes offshore of South Carolina about 145 km due east of Charleston. If they are Late Triassic in age, the distance would be significant because it is equivalent to that between Mt. Kilimanjaro and the Gregory Rift.

Large welt-like uplifts in Virginia, the Highland and Carolina uplifts, have been suggested as being caused by Mesozoic hotspots. Carbonized organic tissues, crystalline 2M illite, and the presence of wurtzite (a high-temperature polymorph of sphalerite) support the postulated Mesozoic hot spot centered at the border between North Carolina and Virginia, with hot springs in the Dan River-Danville basin. The organic tissues found in all of the other basins are lighter in color than those in the Dan River-Danville basin, indicating lower paleotemperatures.

Seismic lines and coreholes through many of the basins show the depths of the remaining tilted deposits. The seismic lines show a base of the Richmond basin at a depth of 1,370 m, and bases of the Culpeper and Deep River basins around 1,500 m. Restoration of the strata to a horizontal position indicates that the thickness of deposits may be 4,600 m in the Dan River-Danville basin.

The sedimentary deposits in the Newark rift system consist of lacustrine shale, siltstone, and sandstone; interior fluvial and valley-floor siltstone, sandstone, and conglomerate; and talus slope sandstone, conglomerate, and fanglomerate. The names of the rock formations are different in each basin, and the deposits in all of the individual basins have been placed by some into the Newark Supergroup.

The events between the abandonment of rifting along the Newark rift system and the opening of the Atlantic are unclear. Radiating tholeiitic dikes cutting across the grain of the Newark rift system have K-Ar dates of Late Jurassic and Early Cretaceous age . Similar K-Ar dates have been measured in alkalic and peralkalic volcanic rocks such as nepheline and analcite syenite, and riebeckite granite of the White Mountain Plutonic-Volcanic Suite which may represent a fossil hot spot in New England.

Post-depositional events have imprinted complex structures over the remnants of the Newark rift system. A recent seismic line that penetrates approximately 24 km in the crust (8 seconds) has been run by L. D. Harris and K. C. Bayer, U.S. Geological Survey, across the Richmond basin in Virginia. It shows the basin to be underlain at depth by several major east-dipping thrust sheets. The tilted basins of the Newark rift system and their dipping rocks appear to be in the process of being thrust out of existence along listric thrusts. Presently active high-angle reverse faulting, and microseismic and macroseismic activity suggest that present-day thrusting is generated by east-west compression. The dip of the thrust sheets shows that the thrusting is coming from the east.

The thrusted nature of the Blue Ridge, Piedmont, Coastal Plain, and Continental Shelf does not allow simple explanations for the relationship between the Newark rift system and the opening of the Atlantic Ocean. The depth of the crust under the Richmond basin suggests that the continental crust was too thick for the continents to separate under that part of the Newark rift system. The compressional stress regime that produced earlier continental collision may have relaxed prior to the evolution of the Newark rift system. Or extension along the Newark rift system may be a result of compression at depth. In the latter scenario, the Newark rift system may have very little relation to the later opening of the Atlantic.

The data are too new to rule out another possible sequence of events. 1) Compression in the Permian resulting in a stacking up of thrust sheets in the Alleghanian orogeny. 2) Extension during the pull-apart stage, which resulted first in the creation of the Newark rift system, and finally in the separation of the North American plate from the Eurasian and African plates at the initial Continental Shelf-Continental Rise boundary that now lies below the present Continental Shelf. 3) Compression from the oceanic basaltic crust after it eventually coupled to the North American plate and formed features that suggest the North American plate is now being pushed by the oceanic crust of the Atlantic Ocean, much like the events forming thrust faults along the east edge of the Red Sea in Saudi Arabia.

The Newark rift system shares many features, such as linear chains of large tectonic lakes, in common with active rift systems. In contrast, it also has features, such as being underlain by a thick continental crust, that may make it unique.

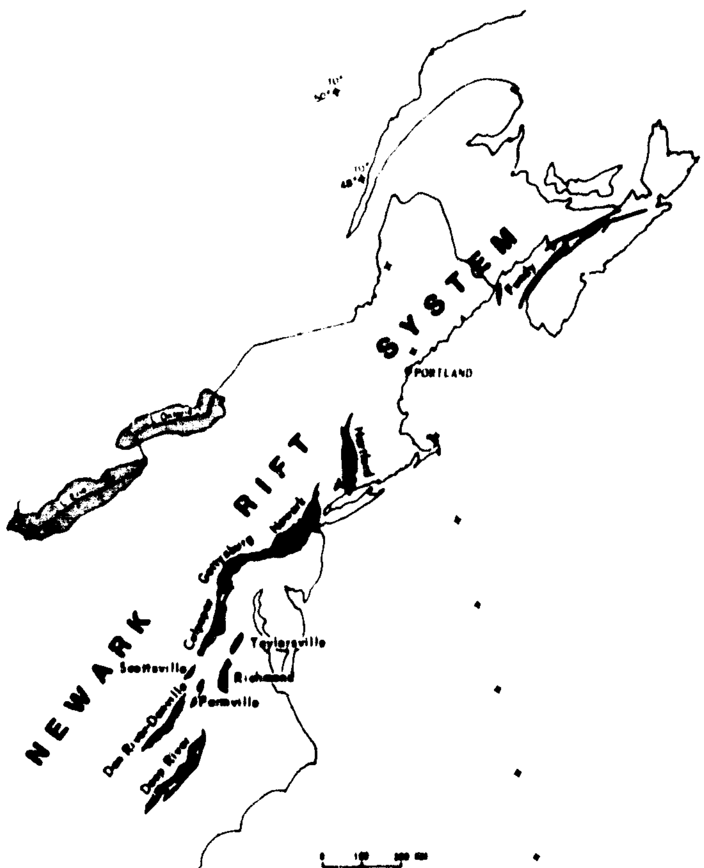

Figure 1. Exposed faults and basins of the Newark rift system.

GEOLOGIC AND GEOCHEMICAL EVIDENCE FOR THE NATURE AND DEVELOPMENT OF THE PRECAMBRIAN MIDCONTINENT RIFT OF NORTH AMERICA, John C. Green, Geology Dept.; Univ. of Minn., Duluth; Duluth, MN 55812.

The Midcontinent rift system of North America (White, 1972; Wold and Hinze, 1982) is about 2300 km long, extending from central Kansas northeastward to Lake Superior and thence southward through Michigan and perhaps into Ohio. Having developed in late Proterozoic time (1200-1100 m.y. ago), it is one of the world's largest as well as the oldest intact continental rift; its characteristics can inform us about such aspects as lithospheric rigidity and the geochemical evolution of the mantle long before "modern" (Mesozoic-Cenozoic) plate tectonics. Although the Grenville Front, with which the Midcontinent Rift is roughly contemporaneous and with which it appears to have a junction (hidden beneath thick Paleozoic sediments) has been interpreted by some workers as a lithospheric suture resulting from plate collision (e.g. Donaldson and Irving, 1972), the Midcontinent structure shows ample evidence of active (heat-generated) rather than passive (collision related) rifting.

Details of the physical and chemical nature of the rift and its products are accessible to us only in the Lake Superior area where these rocks crop out; elsewhere they are buried under Paleozoic strata. Even in the Lake Superior area, the critical axial zone of the rift is hidden beneath thick Upper Proterozoic and Pleistocene sediments and the lake itself. The rift-associated rocks are known stratigraphically as Keweenawan (Morey and Green, 1982). From examination of these around the Lake Superior basin the following history and character of the rift (at least in that area) can be inferred. The general picture that emerges is quite different from the currently most widely described model of continental rifts in which graben structures play a major role.

1. Stratigraphic relations at the base of the Keweenawan lavas do not suggest that there was pre-rift doming or arching, although sensitive evidence is lacking in parts of the region. In fact, the basal lavas in the southwestern portion of the basin conformably overlie mature, cratonic quartz arenite sheets which were contemporaneously being deposited in shallow water when volcanism started. This implies neither an arch nor a graben

2. The principal products of rifting--the immense volumes of Keweenawan lavas (400,000 km^3)(Green, 1977, 1982)--were erupted in broad basins of deposition (lava plateaus) which can now be distinguished by paleomagnetism and stratigraphic mapping aided by geophysics (White, 1972). Individual flows and flow groups of these flood lavas can be traced for tens of kilometers along strike, indicating a level depositional surface. Approximately 8 such plateaus were produced in the Lake Superior area, overlapping to varying extents, as rifting intensity waxed and waned along the rift. Each such plateau subsided centrally during and/or after eruption, and contains 2 to 7 km of lavas in its center; thus each one could be compared to a smaller but thicker Columbia River, Deccan, or Paraná plateau basalt sequence. These flood lavas were evidently fed by fissures; numerous dikes of comparable composition are found cutting both the lavas and the surrounding older rocks (Green, 1977). The great bulk of this volcanism and associated intrusion (e.g. Duluth Complex) occurred in a short time interval (1100 $\pm$ 10 m.y.; Silver and Green, 1972).

3. In all places around the basin, the lavas lap onto older rocks along surfaces of low relief. If normal faulting and graben had been involved, there should be many instances where the lavas are faulted against or abut against older rocks, but nowhere have such relations been found. Neither are normal faults common within the volcanic sequence nor associated with the feeder dikes outside the present limits of volcanic exposures.

4. After volcanism, crustal subsidence continued along the axis of the

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latest volcanic plateaus (Lake Superior syncline), causing the deposition of several km of clastic sediments on top of the volcanic rocks. At intervals during and after this subsidence and sedimentation, reverse faulting and horst development took place (Morey, 1972; Craddock, 1972); these reverse faults are the major structures controlling rock distribution, other than the initial lava basins, in the area. This marked the end of the rift history and the area has been tectonically rather stable for the past 900 m.y. or so, though minor adjustments affecting the thin overlying Paleozoic sediments continued to the south.

5. The lavas and associated intrusions are basically tholeiitic in composition (as in the other major plateau lavas of the world) but the 450 or so analyses currently available show that a wide range of composition is present (Green, 1977; Green et al., unpublished data, in prep.). The initial flows in the earliest plateaus around the basin are more alkaline, yet high in Ni and Cr as well as incompatible elements, but the most abundant types are (a) high-Al olivine tholeiites with undepleted LREE that show strong resemblances to some MORBs. and (b) high-Fe transitional or Fe-Ti tholeiites which resemble the bulk of younger major continental tholeiite provinces. These rocks thus imply a lack of significant mantle evolution in the last 1100 m.y. The most primitive olivine tholeiites have Mg' values near .70 and are similar to direct or only slightly olivine-fractionated partial melts from mantle spinel peridotite. They form one end of a coherent evolutionary trend of iron-enrichment to the high-Fe tholeiites and then alkali-enrichment through basaltic andesites, icelandites, and rhyolites. These magmas must have been very dry, as shown by the rarity of hydrous primary minerals even in highly evolved rocks. Careful modeling of fractional crystallization using both major and trace elements (Brannon et al., 1981) does not give satisfactory results. However, dynamic crystallization experiments at controlled fO_2 and 1 atm (Green, 1980) on a primitive lava sample gave increasingly Fe-enriched residual liquids of similar though not identical compositions to analyzed lavas. The large relative volume of rhyolites in some of the plateaus seems too great to be accounted for by fractional crystallization of mantle melts, but isotopic data imply little crustal contribution (Van Schmus et al., 1982). Pb, Sr, and Nd isotope studies show the mantle source of the lavas to be as much as 4 b.y. old (Leeman, 1977; Dosso et al., 1980). No mantle xenoliths have been found.

The high-Al character of the olivine tholeiites (16-18% Al_2O_3, aphyric) is unique among major plateau basalts, and may be related to the widespread development of anorthosites during the mid- to late Proterozoic in rift environments proposed by Emslie (1978) and others. The associated Keweenawan Duluth Complex, though consisting of cumulate rocks, is plagioclase-rich and a part of the Proterozoic anorthositic suite; its petrogenetic relations to the lavas are under investigation (for initial work see Phinney, 1969).

Several workers (e.g. Chase and Gilmer [C + G], 1973; Weiblen and Morey, 1980) have attempted to interpret the Midcontinent rift in terms of a rigid plate rift separation of up to 90 km, or fault-bounded basins modeled after the East African/Red Sea system, but the geologic evidence from the Lake Superior area shows these models to be deficient on several counts (see also White, 1972). A. Nowhere are Keweenawan lavas seen to be faulted against pre-Keweenawan rocks, which would be expected for a graben or half-graben rift. B. In an actively sinking graben, coarse, immature clastic sediments would be expected to be shed off the adjacent older highlands and to underlie and be interstratified with the lavas. Such sediments are not present beneath the flows except in the Osler Group in northwestern Lake Superior. Interflow

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sediments constitute at most only 2-3% of the lava plateau sequences, and these consist nearly entirely of intrabasin (Keweenawan) clasts, even though transport directions were centripetal (Merk and Jirsa, 1982). Neither have closed-basin sediments such as evaporites been found. C. Dips of the lavas in several plateaus show a decrease with stratigraphic height and actual down-dip thickening of individual units, implying subsidence by warping, not faulting. D. The remarkable lateral continuity of many volcanic units shows a lack of contemporaneous or subsequent faulting. If mafic magma supply to the crustal gravity anomaly were dominated by half-graben faulting producing intrusions (such as the Duluth Complex: Weiblen and Morey, 1980), surely many such faults would have reached the surface through the roof lavas, yet we do not see them. E. Chase and Gilmer's model, which claims crustal separation proportional to the width of the gravity anomalies, ignores the gently-dipping structure of the lava plateaus and does not allow for the existence of the thick Osler and Mamainse Point basalts of Ontario or the 2 to 4 km of basalt beneath the eastern Upper Peninsula of Michigan (Oray et al., 1973). Since the gravity anomalies from which C + G calculated 90 km of crustal separation are due to thick lava basins as well as the feeder dikes they assumed, the actual separation must be much less. F. More recent geophysical studies (Wold and Hinze, 1982) do not support the existence of a great transform fault cutting across Lake Superior, as required by C + G's model. Other proposed transforms fit neither geology nor the geophysical anomalies. G. C + G's rigid-plate model implies simultaneous irruption of mantle magma into the crust over the whole rift, whereas geologic, paleomagnetic and isotopic evidence indicates intermittent activity at different places over a period possibly as long as 100 m.y. (Van Schmus et al., 1982). H. The major faults in the Lake Superior district show reverse, not normal, displacement.

The Midcontinent Rift appears to have been a major active continental rift zone that aborted before actual separation and ocean formation could take place. The nature and amount of the volcanic products most resembles younger flood-basalt plateaus (Parana, Deccan, Karroo) which were associated with nearby complete continental separation and in which broad subsidence rather than graben faulting was dominant. As shown by the great area and volume of mafic rocks in the Lake Superior area (comparable to Iceland) a mantle plume was probably active in this portion of the rift, but partial-melting processes in the mantle both in Archean and Keweenawan times never exhausted ancient, undepleted source rocks.

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ORIGIN AND DEVELOPMENT OF THE OTTAWA GRABEN, S. Kumarapeli,
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The Ottawa graben (Fig. 1), although deeply eroded, displays unmistakable rift valley morphology along a 200 km segment west of Ottawa. This segment was the first to be recognized as a graben - the Ottawa-Bonnéchere graben of Kay (1942). Since then, the graben has been traced northwestwards to Lake Nipissing and Lake Timiskaming areas where the graben bifurcates and its main faults split into divergent smaller faults. Eastwards the graben has been traced into Montreal area and further east its presence beneath the thrust sheets of the Appalachian foldbelt can be inferred from the fact that intrusions localized along graben faults (e.g. Monteregian intrusions) continue well into the foldbelt. Thus, the length of the graben is about 700 km.

The Ottawa graben is superimposed on the North American craton. In places it transects the regional trends of the Canadian Shield. It extends into the continental interior from a prominent salient of the Appalachian foldbelt - the Sutton Mountains salient. The apical area of this salient is characterized by prominent, nearly coincident, positive gravity and magnetic anomalies. Modelling of these anomalies constrained by the geological characteristics of the area indicates that anomalies are the expressions of a thick pile of dominantly mafic, bimodel volcanics, Tibbit Hill volcanics (Fig. 1), which are probably related to the development of the Ottawa graben (Kumarapeli and others 1981). The precise age of these volcanics is not known but they are either early Cambrian or late Precambrian. This volcanism was probably coeval with the emplacement of a prominent dike swarm - the Grenville dike swarm (Fig. 1) - of tholeiitic diabase, along the length of the graben. This episode of mafic volcanism/magmatism was followed by the emplacement of alkaline complexes, including several carbonatite complexes which yield K/Ar ages of about 565 m.y. (Doig 1970). The complexes are concentrated mostly in Lake Nipissing area. A second episode of alkalic magmatism occurred along the eastern part of the graben in the early Cretaceous. The products of this event are the Monteregian intrusions (Fig. 1) which also include two carbonatite complexes.

The time of initial rifting along the Ottawa graben cannot be determined by direct methods. The earliest igneous events - emplacement of the Grenville dike swarm and the volcanism in the Sutton Mountains region - in late Precambrian/early Cambrian times are probably related in time and cause to the initial rifting. The early Cretaceous Monteregian event is probably related to reactivation of the ancient rift zone.

Except for the gravity and magnetic anomalies of the Sutton Mountains region (these are actually located in the Appalachian foldbelt) there are no obvious geophysical anomalies which can be correlated with the Ottawa graben. The fault troughs also do not contain significant amounts of rift-related clastics and/or volcanics. These features are consistent with the great age and the deep erosion of the graben.

Any model incorporating the origin and development of the Ottawa graben must provide a coherent rationale for the following aspects of the graben: (i) The Ottawa graben represents the trace of a tension crack (of the lithosphere) which propagated westwards from the apex of a prominent salient of the Appalachian foldbelt; (ii) The apical area of the salient is underlain by a pile of volcanics which are presumably rift related;

OTTAWA GRABEN

Kumarapeli, S.

(iii) The initial rifting and the beginning of the Appalachian stratigraphic sequences are nearly contemporaneous (note that the Tibbit Hill volcanics occur at the base of the Appalachian stratigraphic sequences in the Sutton Mountains area).

Dewey and Kidd (1974) have suggested that the Iapetus Ocean, whose telescoped scar represents the internal zones of the Appalachian orogen originated by continental separation in latest Precambrian - earliest Cambrian time following late Precambrian continental distension and the formation of a multibranched rift system, the eo-Appalachian rift system (Kumarapeli, 1976). The Tibbit Hill volcanics and other rift facies volcanics, contained in early stratigraphic sequences of the Appalachian foldbelt (Rankin 1976), are probably related to this early rifting. One of the rrr triple junctions of the eo-Appalachian rift system was located in the Sutton Mountains region. Two arms of the Sutton Mountains triple junction apparently went through a Wilson cycle whereas the other arm continued as an aulacogen, which is the Ottawa graben. During the early stages of the evolution of the Iapetus, the paleogeography and the tectonic setting of the Sutton Mountains area may have resembled those of the Afar triple junction, the relationship of the Ottawa graben to the early Iapetus being similar to that of the East African rift system to the Red Sea and the Gulf of Aden. Thus the Tibbit Hill volcanics formed in an environment similar to that of the Afar triangle. Coeval to this volcanism was the emplacement of the Grenville dike swarm whose

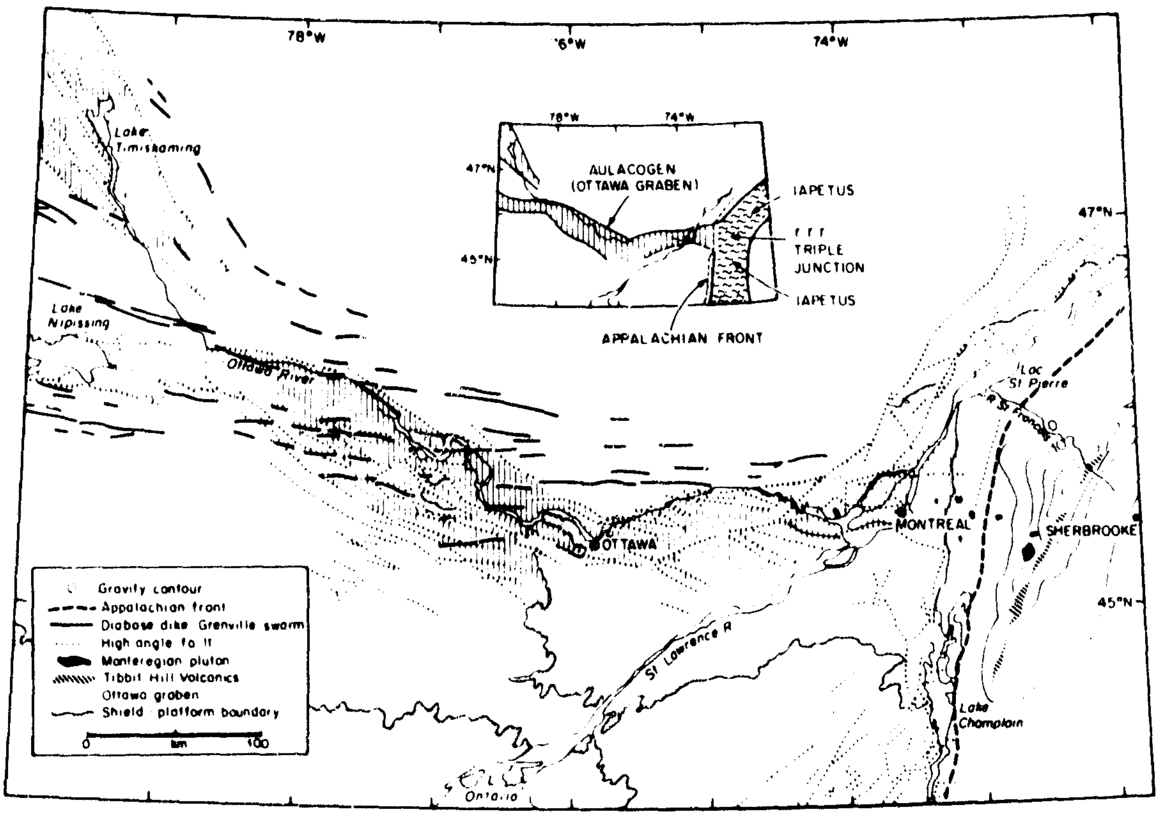

Fig. 1. Key geological elements of the Ottawa graben area. Inset, interpretive sketch of the postulated Sutton Mountains rrr triple junction.

trend closely follows that of the Ottawa graben (Fig. 1). By the late Cambrian time deep erosion had stripped the graben of much of its sedimentary - volcanic fill. The erosion was followed by a shallow marine transgression and the rifted area became buried under an early Paleozoic platformal cover which in turn has been stripped from parts of the graben exposing the Precambrian basement. The early Cretaceous reactivation of the graben and the emplacement of the Monteregian plutons may have been related to the opening of the Atlantic Ocean.

The carbonatite complexes related to both magmatic events are Nb-RE bearing. The ore deposits occur in the form of pyrochlore disseminations in carbonatite host-rock.

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TIMING AND CHEMISTRY OF IGNEOUS EVENTS ASSOCIATED WITH THE SOUTHERN OKLAHOMA AULACOGEN: M. Charles Gilbert, Dept. of Geological Sciences, Virginia Polytechnic Institute and State University, Blacksburg, Va. 24061.

Introduction: The fundamental geologic setting of the Anadarko-Ardmore basins and Arbuckle-Wichita uplifts was not fully appreciated until the work of Ham, Denison, and Merritt (1964). Igneous activity and/or large movements have characterized the region from mid-Proterozoic through Permian, much of it involving rift-like features. Subsequently, an attempt to relate this setting to modern plate tectonic theory was made by Hoffman, Dewey and Burke (1974), who popularized the designation "aulacogen" for the site. The area is important to understanding the growth, stabilization, chemistry, petrology, and tectonics of the North American craton. The Anadarko Basin is the deepest, large basin in North America with a full 12-15 km of in-place, vertical section. It is bounded on the south by the Wichita uplift fault zone where there is at least 10 km of throw. The earliest igneous events are only poorly constrained, but a large, well-dated rhyolitic episode occurred during Middle Cambrian. While the general chemistry and petrography are matched in much of the Mid-Continent basement, as can be seen at the surface in the St. Francois Mtns., Mo., Wolf River batholith, Wisc., and Pikes Peak sequence, Colo., the Wichita ages are completely anomalous, being roughly 1/2 to 1 by younger. The area of Cambrian age silicic activity is about 350 km long by 40-100 km wide, represents about 40,000 km^3 of magma, and is also anomalous for the large proportion of extrusives to intrusives, 10:1 (Gilbert, 1978). A COCORP line, in two parts totalling 200 km, has just been completed across the Hollis Basin - Wichita Mtns. - Anadarko Basin (Brewer and others, 1981). These data modify some of the structural style previously accepted for the region but none of the perspective outline above.

Sequence of events: The evolution of the region is illustrated in Fig. 1. The earliest evidence of rifting is that of Brewer and others (1981) who suggest that the Pennsylvanian Burch Fault had a prehistory dating back to the Proterozoic because it appears to terminate a large, previously unknown, basin filled with about 10 km of presumed sediments. The sense of movement at that time is unknown, but these workers preferred north side down, opposite to what is shown in the figure. They also believe that igneous activity at 1.35-1.4 by in the Arbuckle Mtns. may be an event related to the faulting.

Two episodes of basaltic intrusion occurred which are not yet reliably dated (Powell, Gilbert and Fischer, 1980). The earlier phase was tholeiitic, forming a Layered Complex (15 x 65 km min. size). This body was subsequently cut by an unknown number of biotite gabbro bodies that are layered and locally differentiated. The largest known is 8 km in diameter but they occur over a distance of 155 km along the length of the present uplift and may, in fact, be volumetrically more significant than the Layered Complex. Powell (1979) has argued that this magma was a hydrous tholeiite rather than one with alkaline affinities (Table 1). Ham, Denison, and Merritt (1964) assumed that the spilitic basalts of the subsurface (Navajoe Mountain Basalt Group) were the extrusive equivalent of the Layered Complex gabbros, but the two distinct phases of gabbroic intrusives were not appreciated at the time.

After a period of uplift when several kilometers of overburden were stripped, a period of intense, Mid-Cambrian rhyolitic volcanism occurred. Ham, Denison, and Merritt (1964) argued for faulting (rifting) just preceeding or accompanying eruptions based on relations in the Arbuckles. Eruptive style may have been unique for the Earth since no calderas have been identified. The high ratio of extrusives to intrusives is consistent with lack of calderas. This style may have been caused by 1) the particularly dense substrate of gabbro not allowing ponding of rhyolitic liquid in the upper crust, 2) the estimated high temperatures of the rhyolites (>900-1000° C), and the very dry (estimated wt% H_2O < 1) character of the liquid (Gilbert, 1978). Chemistry of the granite sills directly related to the volcanics, is one of extreme depletion in some elements such as MgO and CaO. This argues for either a high degree of differentiation or remelting of a previously melted and drained lower crustal section (Table 1).

Because of substantial post-consolidation recrystallization in the rhyolites, the chemistry of the silicic magma is best discerned from the equivalent granites. Myers and Gilbert (1980) found that the granite chemistries matched well with the criteria set out by Loiselle and Wones (1979) for A-types. They further found that Wichita granites could be grouped into 3 chemical classes. For example, Mt. Scott Granite (and class) has an extremely uniform concentration of Sr over a lateral distance of 50 km probably arguing for an origin by fractional crystallization.

The last, documented igneous event in the Wichita block is a set of diabase dikes. Generation of 40,000 km^3 of rhyolitic liquid requires at least as much basaltic liquid, either as the heat source for partial melting or as the chemical source on fractionation. Presumably, the late diabases, which are olivine normative, are our record of this liquid (Table 1).

After cessation of rhyolitic volcanism, the region sank and received 3 to 5 km of sediments through the Mississippian, with a large proportion being shelf-type carbonates. Episodic uplift of the Wichita block occurred throughout the Pennsylvanian, with concomitant sinking of the Anadarko Basin. Eroded rock from the Wichita block was shed off to the sides of the uplift, with 4 to 5 km of Pennsylvanian section generated in the Anadarko. Two to three km of additional Permian fill came from the Ouachitas to the southeast and finally buried the Wichitas. The linked uplift and sinking was accomplished by crustal rupture with faults of large throw, which have been interpreted as high-angle reverse and also as major left-lateral, strike-slip. It appears that the Cambrian extrusives, the upper Paleozoic depositional basins, and the Pennsylvanian fault systems have a basic en echelon pattern to, and lie athwart, the regional structural highs.

The history of the aulacogen is well-determined from the Mid-Cambrian forward. Problems which must be resolved before the earlier history, as depicted in Fig. 1, is well constrained include: a) age of basaltic intrusion #1 (Glen Mountains Layered Complex) and #2 (Roosevelt Gabbros), b) age, nature, and extent of the Proterozoic (?) pre-Burch fault, and c) distribution, age and character of the Proterozoic sediments south of the Wichitas. None of the uplifts or sinkings can be accounted for simply by melting and emplacement. Regional thermal events or lithospheric stress systems considerably larger in size than the distribution of the igneous bodies is necessary.

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Table 1. Chemistry of Selected Igneous Units Wichita Mountains, Oklahoma.

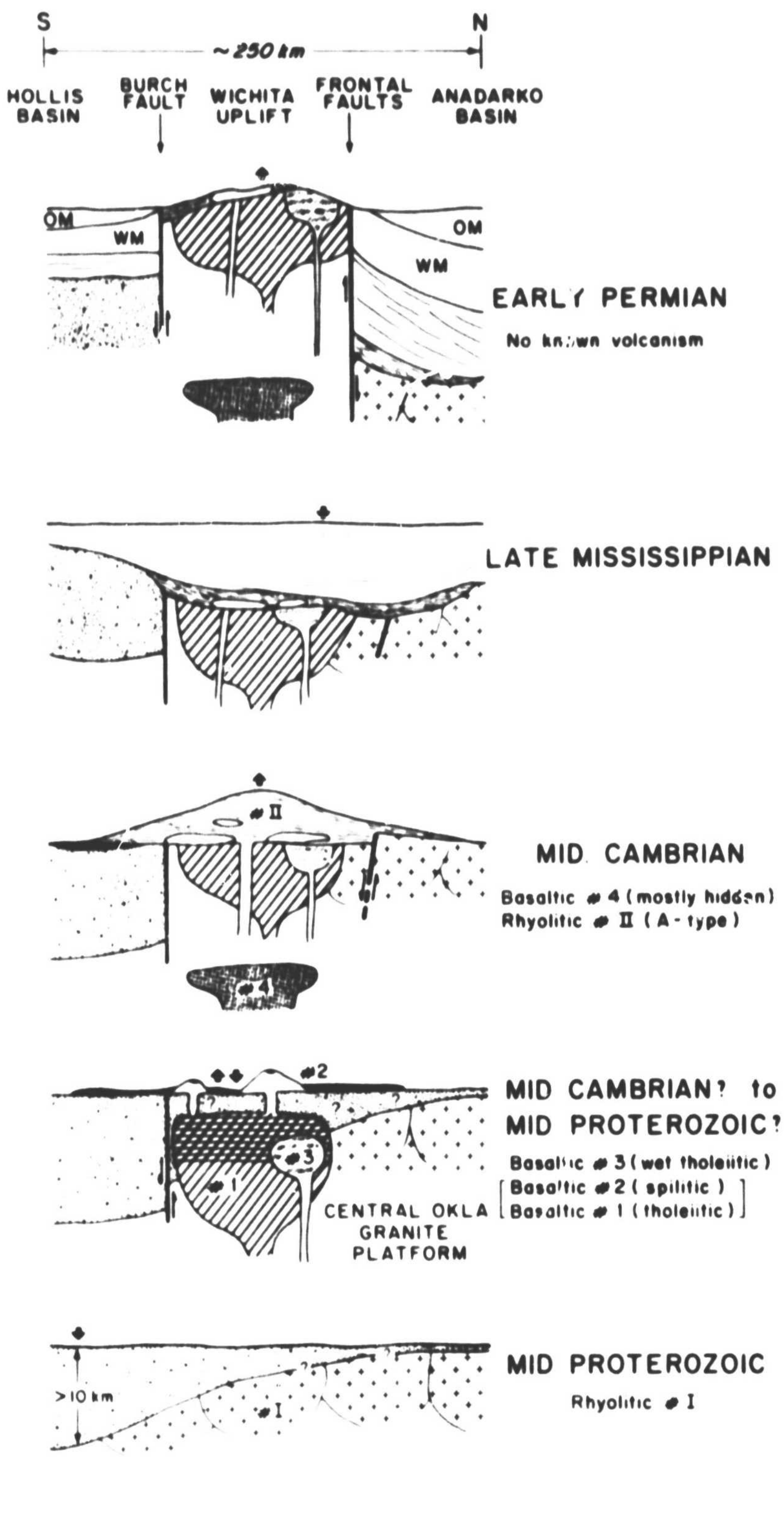
Wt.%	(Pre-granite) Roosevelt Gabbros[1]	(Post-granite) Late Diabase[2]	(Early-granite) Mt. Scott Granite[3]	(Late-granite) Quanah Granite[3]
SiO_2	47.3	46.6	72.3	76.2
TiO_2	3.0	3.6	.44	.16
Al_2O_3	15.0	13.5	12.3	11.8
Fe_2O_3	13.5	16.4	3.9	2.4
MnO	.21	.21	.08	.02
MgO	8.2	5.4	.31	.03
CaO	8.8	8.7	1.2	.23
Na_2O	1.99	2.3	3.8	4.0
K_2O	.36	.75	4.3	4.75
P_2O_5	.22	.61	.08	.01
Sr ppm[2,3]	ND	371	91	9
Rb ppm	20	22	127	169

1) Average from Table 6, Powell, Gilbert and Fischer, 1980. All Fe as Fe_2O_3.
2) Average from Gilbert and Myers, in preparation.
3) Average from Myers, Gilbert, and Loiselle, 1981.

Timing and Chemistry ... Southern Oklahoma Aulacogen

Gilbert, M. C.

FIG. 1 HISTORICAL DEVELOPMENT
SOUTHERN OKLAHOMA AULACOGEN



This diagram is drawn across the present Wichita Mountains. Short arrows record some of the major net vertical crustal motions necessary during the time frame indicated. Motions shown on faults are not meant to imply lack of horizontal offset.

Rhyolite episodes: #I - shallow granites of 1.2 - 1.5 by apparently constitute the crustal framework; #II - includes both the Wichita Granite Group and the Carlton Rhyolite Group.

Basaltic events: #1 - Glen Mountains Layered Complex; #2 - Navajoe Mountain Basalt - Spilite Group; #3 - Roosevelt Gabbros; #4 - Post-granite, pre-Reagan ss (base of Paleozoic sedimentary section overlying Carlton Rhyolite). Only represented at surface by diabasic dikes.

Top diagram: WM - erosional debris from the Wichita block; OM - erosional debris from the Ouachitas which is "post-tectonic" and eventually buries the uplift.

PULSATION TECTONICS AND RIFTING OF CONTINENTAL MARGINS by Robert E. Sheridan,
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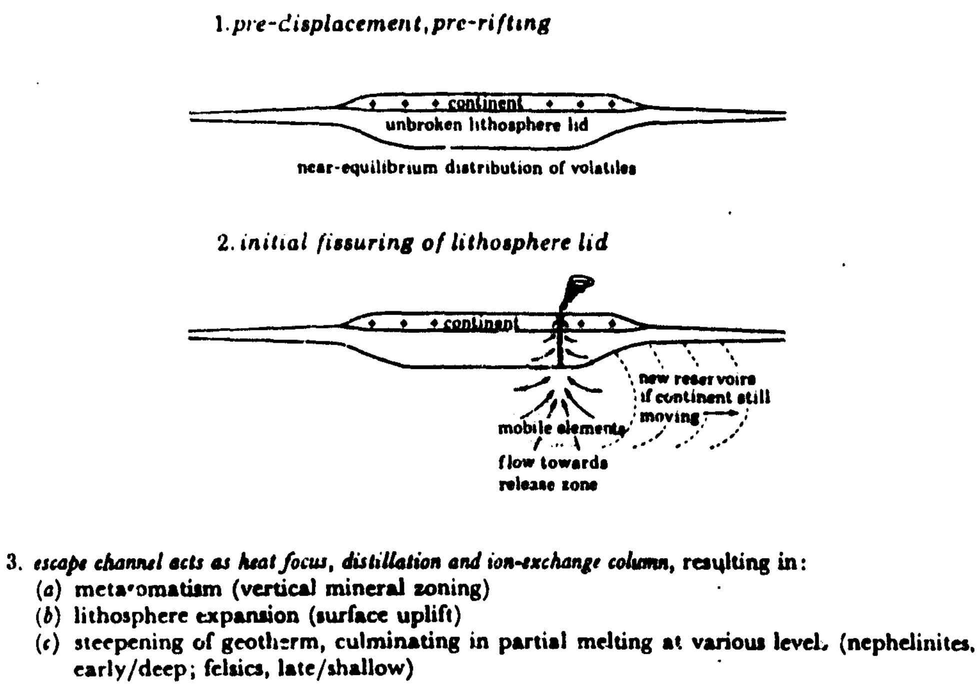
New data from the recent IPOD drilling of DSDP Site 534 in the Blake-Bahama Basin give a definitive age for the spreading-center shift involved in the early rifting of the North American Atlantic margin. A basal Callovian age (~155 m.y.) is determined for the Blake Spur magnetic anomaly marking this spreading-center shift that signals the birth of the modern North Atlantic Ocean. This is some 20 m.y. younger than previously thought. One implication of this surprising result is that this spreading-center shift starting North Atlantic rifting is now of an age which could be assigned to the spreading-center shift needed to end the rifting in the Gulf of Mexico. It is suggested that this might be one and the same event. Another implication of this surprisingly young age for the Blake Spur event is that very high spreading rates are now required for the Jurassic outer magnetic quiet zone along the North American margin. This association of a high spreading rate with a magnetic quiet zone is similar to that for the mid-Cretaceous and implies a link between the processes controlling plate spreading, which are in the upper mantle, and the processes controlling the magnetic field, which are in the outer core. A theory of pulsation tectonics involving the cyclic eruption of plumes of hot mantle material from the lowermost mantle could explain the correlation. Plumes carrying heat away from the core/mantle boundary later reach the asthenosphere and lithosphere to induce faster spreading. The pulse of fast spreading in the Jurassic apparently caused the rifting of the North Atlantic. Other pulses of fast spreading appear to correlate with major ocean openings on various parts of the globe, implying that this might be a consistent process. Rifting of passive margins may be controlled by the more fundamental global processes described by the theory of pulsation tectonics.

MAGMATIC EVIDENCE ON THE CHEMICAL AND THERMAL EVOLUTION OF CONTINENTAL RIFTS: D. K. Bailey, Dept. of Geology, U. of Reading, England

Any concept of the development of continental rifts requires the integration of observations on lithosphere structure and horizontal movement, vertical movements, and magmatic geology, with the more customary geologic and geophysical evidence. Timing is crucial, and only the magmatism can chronicle the chemical and thermal changes in the underlying mantle. Control of the activity (and hence the rifting) by older structures in the lithosphere is clear from the repetition of alkaline magmatism, which marks the release of gas-charged magmas from deep sources repeatedly tapped by reopening of lesions in the continental plate (Bailey, 1977). Frequently this activity is found in ancient cratons where it perforates deeply eroded sections of granulites, themselves highly depleted in volatiles and presumably indicating a similar condition in the underlying mantle. Antiquity of the lithosphere, its great thickness, and the chemistry of the magmatism, rule out recycling of volatiles from the outer part of the Earth to the deep interior (Bailey, 1980a). A rapidly growing body of observation and opinion is in favour of "enrichment" of the mantle to explain alkaline magmatism (Bailey, in press). Often the argument is based on the high levels of incompatible elements that are inexplicable by melt fractionation and must be attributed to source characteristics. Hence the rift enigma — highly enriched magmas from supposedly depleted sources.

Sub-aerial rifting of cold lithosphere would allow the initiation and establishment of effectively continuous degassing, draining a large reservoir in the mantle through the relatively narrow rift zone (Fig. 1). Funnelling of volatiles through the rift zone focusses heat and builds up the level of incompatible elements.



(Figure published courtesy of Phil. Trans. Roy. Soc. London)

FIGURE 1. Schematic diagram showing fissuring of the lithosphere forming an open system through which mantle volatiles migrate to the surface. (Taken from Bailey, 1980a).

With decreasing pressure the rising fluids must cause precipitation of incompatible elements (metasomatism) in the cooler channelways in the overlying mantle and crust (Fig. 2). Ultimately the heating of the channelways and the lowering of the solidus will culminate in melting. This neatly solves two of the chief problems of rift magmatism — its strict localisation and the special chemistry of the magmas. By metasomatic expansion long-lived uplifts of the crust can develop, and these will survive until prolonged heating of the rift segment causes collapse by thermal decomposition of the low density minerals (Bailey, 1972; 1978). The mantle exhaust system thus functions in an exactly opposite way to that postulated for convective plumes (narrow rising column, spreading near the surface). Low initial concentrations of volatiles in a large mantle reservoir are focussed and concentrated in the comparatively small volume of the rift zone.

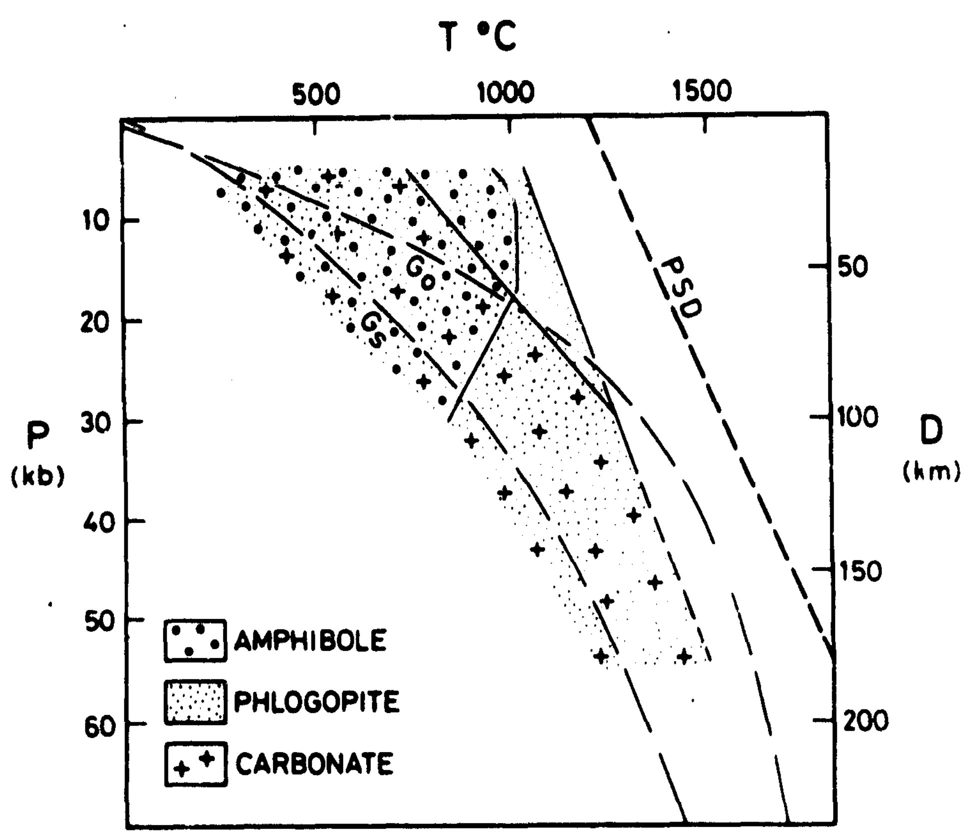


FIGURE 2. Stability fields of possible metasomatic minerals in the mantle, with shield (G_s) and ocean (G_o) geotherms. PSD is the vapour absent peridotite solidus. Experimental sources: Kushiro, 1970; Yoder and Kushiro, 1969; Eggler and Holloway, 1977; Wyllie, 1977. Taken from Bailey, 1979 and in press.

Chemical and thermal evolution

Bailey, D. K.

By combining geochemical, petrological, and experimental evidence it can be shown that diamondiferous kimberlite activity is the limiting case of cratonic magmatism triggered by gas fluxing (Curve S, Fig. 3). Other types of alkaline and carbonatite magmatism are then related to gas fluxing along steeper geotherms, such as Curve G. Melt diapirs may form, and surface eruption of high temperature liquids is possible (Fig. 3). Melts from steeper geotherms may rise at rates slow enough to permit re-equilibration, fractionation, and reaction.

In any area the magmatism will express the interplay of the pre-existing compositional complexity, thermal structure and volatile activity. In the interior of a stable continental plate the magmatic expression may take the following chief forms, either separately, or sequentially if a gas-flux/heating system becomes established. Impersistent cracking of craton nuclei alone permits only kimberlite release, while establishment of a rift transect (West Rift, E. Africa) allows potassic ultramafic melt eruption. With steeper geotherms, as around craton margins, the whole gamut of alkaline magmatism becomes possible (East Rift, E. Africa). But in young lithosphere, marked by more recent tectonic and igneous activity (and with a correspondingly complex thermal structure and compositional heterogeneity) quite different kinds of magmatism may result, especially where pre-existing hydrous minerals can influence the melt generation. Effects of syngenetic H_2O released in the higher parts of the rift segment may be expected to be profound, permitting an even wider spectrum of eruptive lavas as exemplified in the Rio Grande Rift.

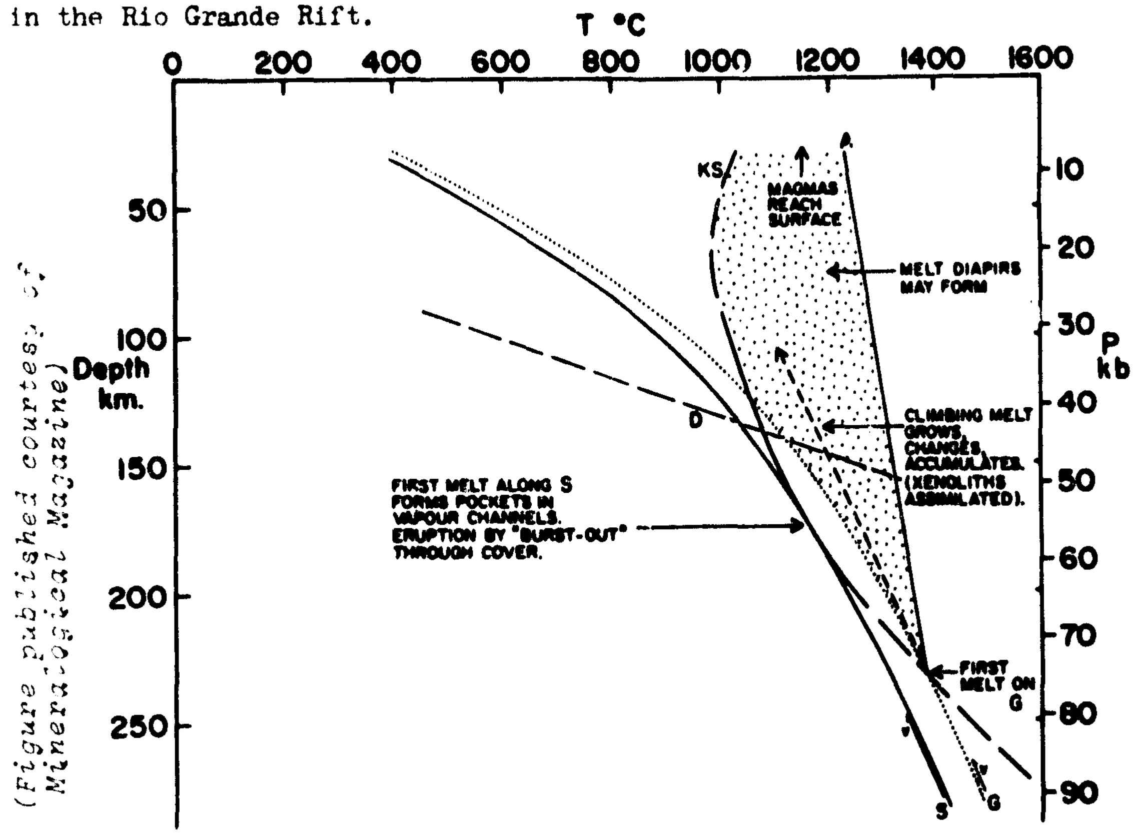


FIGURE 3. Contrasting types of magmatism that would result from volatile (V) uprise through two lithosphere segments with geothermal gradients S and G. A is the melt adiabat; D is diamond stability boundary; KS is kimberlite solidus (Eggler & Wendlandt 1978). From Bailey, 1980b.

Chemical and thermal evolution

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EXPERIMENTAL PETROLOGY AS A PROBE OF RIFTING PROCESSES
By R. F. Wendlandt

A fundamental mechanism of some continental rifting processes appears to be the transmission of thermal energy into the lithosphere by asthenospheric upwelling (the "active" mechanism of Sengor and Burke, 1978). The abundance and variety of rift related magma types which derive from the thermal perturbation suggest that studies of igneous petrogenesis may offer a method of investigating the physical as well as the chemical conditions of rift-forming processes. Systematic variations in the temporal-spatial-compositional relations of igneous rocks are becoming well documented for a number of different rifts (Mohr, 1970; Gass, 1970; Lipman, 1969; Duncan et al., 1972; Lippard and Truckle, 1978; Norry et al., 1980; Brooks and Rucklidge, 1974; Williams, 1972; Baker et al., 1972). It is generally observed that the degree of silica-saturation of eruptive lavas within rifts increases with time while incompatible element contents decrease. Furthermore, at any given instant, lavas erupted within rifts are less silica-undersaturated than lavas erupted outside the rift. These observations are consistent with decreasing depths of origin of magmas with time as would be predicted for magmagenesis associated with an upwelling source region. Accordingly, if the physical and/or chemical conditions of origin of the various magmatic products can be deduced, then it may be possible to constrain mantle plume dynamics and rift processes. In this paper, 3 active-type rifting events which occurred in Africa during periods of slight plate motion (Briden and Gass, 1974; Burke and Wilson, 1972) will be considered.

Petrogenetic Grid

Table I summarizes the conditions of origin of some principal magma types, especially alkaline varieties, associated with continental rifts, as determined by experiments in both simple and natural systems. It is notable that the pressures of origin span the entire lithosphere ranging from depths in excess of 170 km to crustal depths. Also, the temperatures of origin, particularly for the alkaline partial melts, tend to cluster around 1200°C ± 100°.

Partial melting in the upper mantle is likely to occur in the presence of small amounts (<1%) of volatiles (H_2O, CO_2, CH_4, etc.). A consequence of small amounts of volatiles in the source region is that the temperature of beginning of melting and the initial melt composition will be controlled by a melting reaction involving a volatile-bearing mineral (carbonate and/or phlogopite at $P > 25$ kbar; amphibole and/or phlogopite at $P < 25$ kbar; and, amphibole, phlogopite, or carbonate at $P \approx 25$ kbar) and a buffered vapor composition (Eggler 1977, 1978; Wyllie 1978, 1979). Because the vapor is buffered at different compositions with changing P and T, the initial melt compositions change systemmatically with P and T. These controls on partial melting facilitate determining the conditions of origin of a particular magma (of course, the uncertainty of the primary nature of the magma remains). Predicted conditions of formation will have a higher degree of reliability if the magma in question was generated near the solidus by small degrees of melting; with increasing degrees of melting, the initial distinct alkaline character of a partial melt is diluted by the addition of the more refractory components. In the following discussion, emphasis will be on the physical and chemical conditions of origin of primitive alkalic varieties of magmas and the flood sequences, inasmuch as it is believed that these occurrences are minimally evolved. Barberi and Varet (1975) have shown that Ethiopian fissure basalts are less fractionated than basalts from central volcanoes.

EXPERIMENTAL PETROLOGY OF RIFTS

Wendlandt, R. F.

Magmagenesis and Constraints on Rift Processes

Figures 1, 2, and 3 are sections across the Ethiopian and E. African rifts, and across the Angolan alkaline igneous province, respectively, showing depths of origin of the eruptives. For the E. African and Ethiopian occurrences, there are sufficient numbers of ages to show the temporal evolution of the subcrustal mantle. The 1200°C continental isotherm (Pollack and Chapman, 1977), corresponding to a heat flow of 40 mWm^{-2}, is shown in all the profiles at approximately 180 km depth. Because many of the alkalic magmas are generated at about 1200°C (Table 1), the asthenospheric upwelling, outlined by the depths of origin of the magmas, is also approximately defining the encroachment of the 1200°C isotherm upwards. The series of sections may, thus, be interpreted as depicting the ascent of an asthenospheric thermal plume with concomitant thinning of the lithosphere.

The lateral dimension of the area affected by the asthenospheric upwelling is variable in the 3 rift sections; up to 800 km in Ethiopia, 300 km in Kenya but approximately 1000 km if both East and West Rifts are included, and approximately 1000 km in Angola (a half-rift). The former 2 regions are characterized by steep eastern and western shoulders and assymetric northern and southern shoulders on the anomalous upwelling (an observation in concert with teleseismic data by Long, 1976, in NW Kenya) and by assymetrical dispositions of rift relative to upwelling asthenosphere. The Recent section (D) of depths of origin across E. Africa is very similar to sections of Girdler et al. (1969) and Baker and Wohlenberg (1971) based on gravity measurements, Long and Backhouse (1976) based on teleseismic data, and contours of current lithospheric thickness in Africa (Fairhead and Reeves, 1977) based on teleseismic and gravity data. The Angolan section shows a transitional descent to depth away from the S. Atlantic margin. Several different interpretations of the differences between the Angolan and E. African settings are possible. The Angolan volcanics may represent a plume trace on a moving plate: although this is not consistent with Briden and Gass' (1974) estimate that the African plate was approximately stationary prior to the opening of the S. Atlantic, additional ages on the eruptives are needed to constrain the observation. Alternatively, the Angolan rift may be different from the other rifts in that it was a successful rift, leading to continental separation and highly attenuated continental crust, whereas the Ethiopian and Kenyan rifts have not succeded. Gass et al. (1978) suggested that Mesozoic thermal perturbations associated with the breakup of Gondwanaland were of a greater magnitude than Cenozoic hotspots.

Depth estimates to anomalous mantle that are based on seismic studies are shallower than minimum depths of origin of most recent rift eruptives estimated from experimental studies. Seismically anomalous material at shallow depths in rifts may be dike injection zones as suggested by Baker and Wohlenberg (1971) and Fairhead (1976) and not melt source regions.

The diversity of depths of origin of the rock types in E. Africa, ranging from 170km to the lower crust, combined with an abundance of dated occurrences permits calculation of the rate of ascent of the asthenosphere (Fig. 4). This calculation is based on the most recent diamond-bearing kimberlite age, the oldest carbonatite age in the E. Rift, and the ages of initiation of eruption of flood basalts, phonolites, and trachytes. A systematic decrease in the rate of ascent of asthenosphere is notable, and provides a basis for future thermal modelling of plume dynamics.

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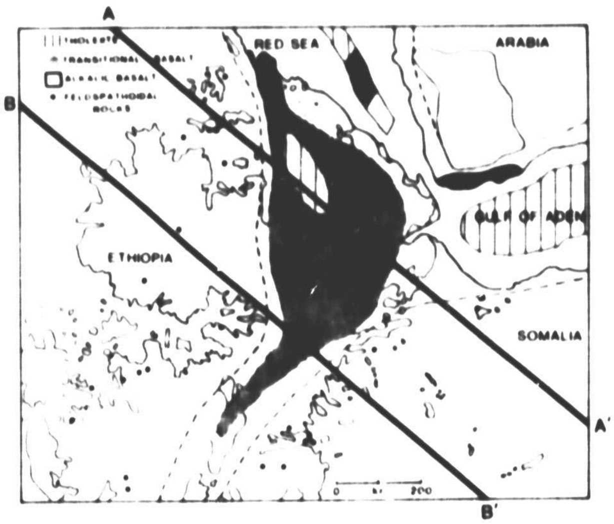

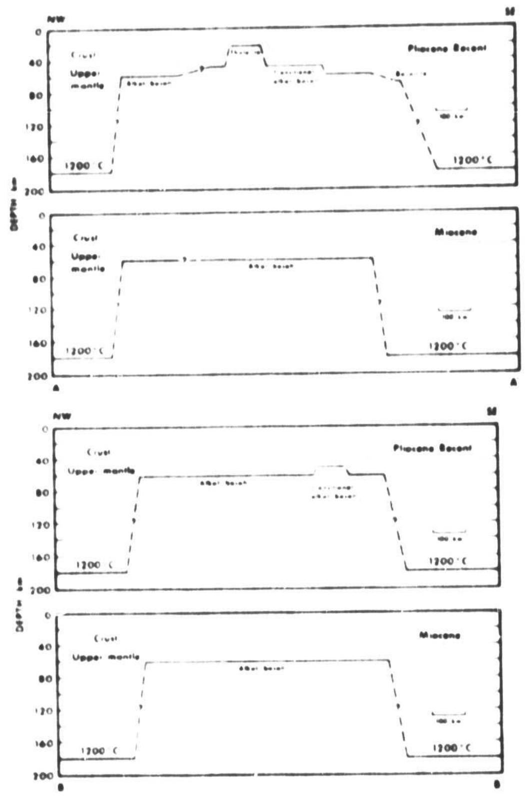

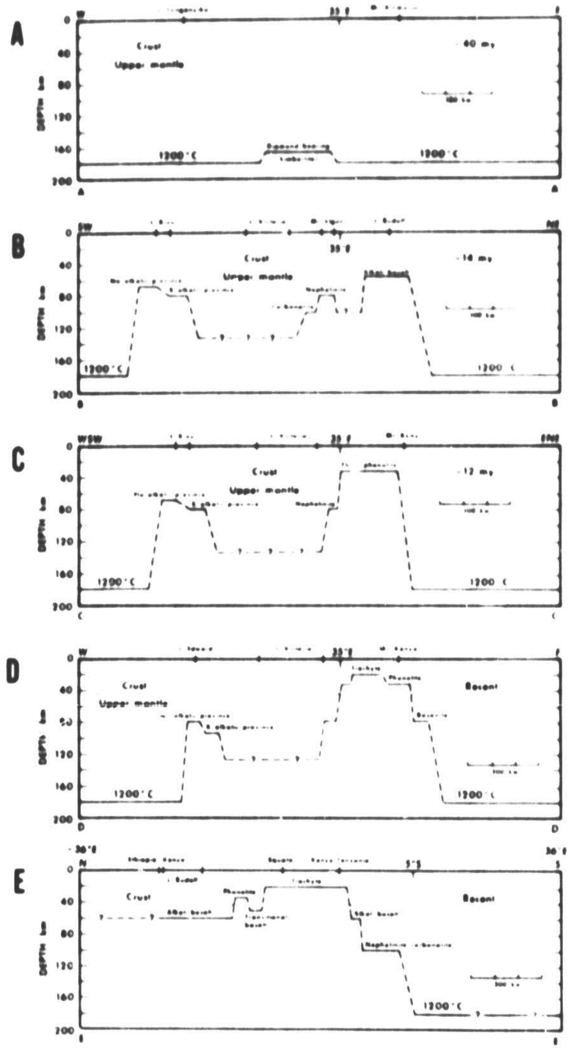

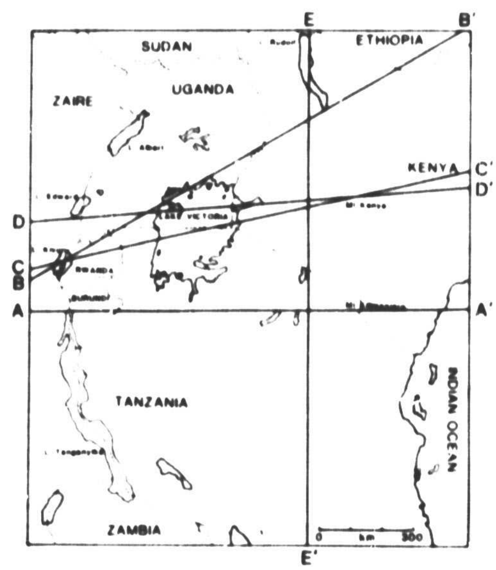

Fig. 1 (above)
Sections across the Ethiopian rift showing depths of origin of volcanics. Section locations are shown on the map. Temporal and spatial relations of volcanics are from LeBas and Mohr (1968), Gass (1970), Mohr (1970, 1971), Baker et al. (1972), Cox et al. (1973), Gass et al. (1973), Barberi and Varet (1978), and Shakhov (1976).

Fig. 2 (right and upper right)
Sections across the E. African rifts showing depths of origin of volcanics. Section locations are summarized on the map. Temporal and spatial relations of volcanics are from Williams (1939), Edwards and Howkins (1966), Dawson (1970), King (1970), Williams (1970, 1978), Baker et al. (1971), King and Chapman (1972), King et al. (1972), Nixon (1973), Davis (1978), Lippard and Truckle (1978), and Pouclet (1980).

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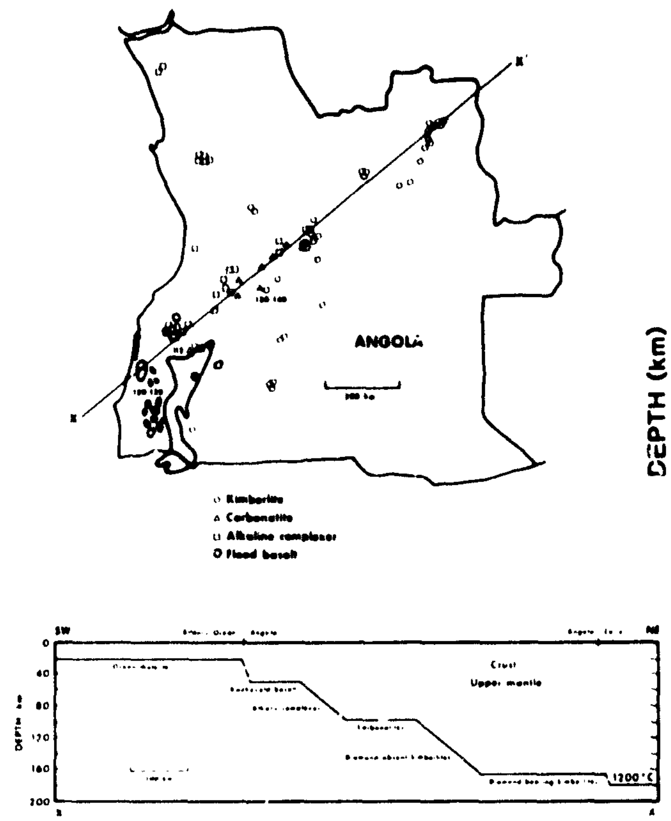

Fig. 3 A section across Angola showing depths of origin of volcanics. Temporal and spatial relations of volcanics used in the construction are from Sousa Machado (1958), Siedner and Miller (1968), Lapido-Loureiro (1968), Reis (1972), Allsopp and Kramers (1977), Davis (1978), and Macintyre (1978).

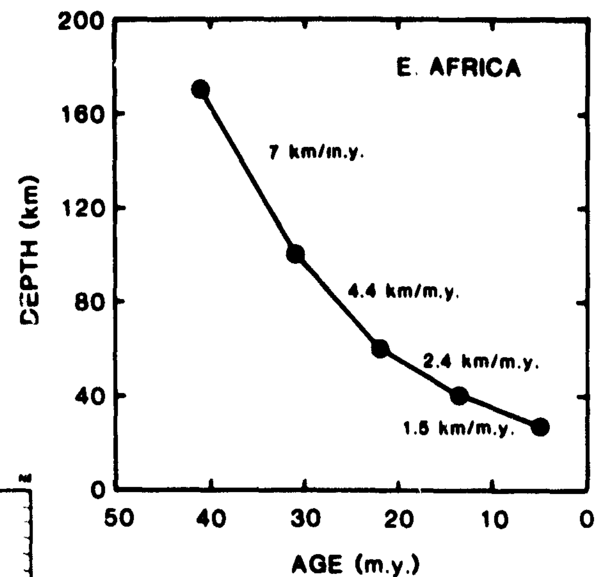

Fig. 4 Calculated ascent rate of asthenosphere under the E. Rift in Kenya-Tanzania.

Table 1 Conditions of Origin of Magmas Associated with Rifts

Magma Type	Conditions of Origin P (kbar)	Conditions of Origin T (°C)	References
Ocean Tholeiite	8-10		Kushiro (1973), Hodges and Bender (1976), Fujii and Kushiro (1977), Bender et al (1978), Presnall et al (1978)
Olivine Basalt			
Hawaiian Olivine Tholeiite	9-13.5		Green and Ringwood (1967), Green (1970)
Karroo Olivine Basalt	12		Cox and Jamieson (1974)
Transitional Alkali Basalt	15		Green (1970)
Alkali Olivine Basalt	10-15+	1225-1325	Jaques and Green (1980)
Skye	17	1380	Thompson (1974)
Basanite	18-24 15-21		Bultitude and Green (1971) Green (1970)
Sodic Alkalic Magmas			
Phonolite	10-15	1100-1150	Kushiro (1968), Irving and Price (1981)
Nephelinite	13-22	1070-1210	Olafsson (1980)
Olivine-Melilite-Nephelinite			Green (1970)
Olivine-Nephelinite	15-30		Green (1970)
Olivine-rich Basanite			Green (1970)
Olivine Melilitite	27	1150-1200	Brey and Green (1977)
Potassic Alkalic Magmas			
Olivine normative	14-17	1100-1150	Wendlandt and Eggler (1980b)
Transitional Alkaline	15	1100-1150	Wendlandt and Eggler (1980b)
Leucite normative	17-25	1125-1200	Wendlandt and Eggler (1980b)
Kalsilite normative	25-30	1150-1200	Wendlandt and Eggler (1980b)
Carbonatite	25-30	1150-1250	Wyllie and Huang (1975 a,b, 1976), Eggler (1975, 1976, 1978) Wendlandt and Mysen (1980), Wendlandt and Eggler (1980 a,b)
Kimberlite			
Diamond-absent	<30	<1150	Wendlandt and Eggler (1980b), Eggler and Wendlandt (1979)
Diamond-bearing	47-75	1100-1300	Kennedy and Kennedy (1976), Wyllie (1980)

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PETROGENESIS OF SILICA-SATURATED EVOLVED RIFT MAGMATIC ROCKS

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Continental rift magmatic products vary greatly in petrological character and in many rifts include large volumes of silicic rocks. MacDonald (1974) has, on the basis of volcanic associations and chemistry, suggested that these rocks may be grouped into three main types of petrogenetic series:

§1. Olivine melanephelinite-nephelinite-phonolite (+ carbonatite).

§2. Approximately silica-saturated alkali basalt (or olivine tholeiite)-mugearite-trachyte < comedite or pantellerite / trachyphonolite-phonolite

§3. Quartz tholeiitic or calc-alkaline series.

Series §1 and §2 are by far the most common ones, and are described from rifts of very different ages, e.g. the Precambrian Gardar rift, South Greenland (Upton 1974), the Permian Oslo rift, South Norway (Barth 1945), and the still active East African and Rio Grande rift systems (Baker et al. 1978; Baldridge 1978). Approximately silica-saturated magmatic rock suites (§2) are also described from a number of oceanic islands (Macdonald 1974). Studies of the conditions of formation of the different members of the rock series §1 and §2 are therefore essential to our understanding of rift magmatism in general, and may also increase our understanding of magmatism in other tectonic settings.

I have studied trend §2 in the Permian Oslo paleorift for two reasons:

1. The students of young rifts can observe only the uppermost features of rift systems, whereas the rocks in the Permian age Oslo rift are more deeply eroded and thus both volcanic rocks and their plutonic associates are exposed.
2. The complete tectonomagmatic history may be considered for a paleo-rift, whereas an active rift reveals only the earlier stages of its evolutionary history.

Magmatic activity in the Oslo rift started about 296 Ma ago with emplacement of a few felsic sills, followed by fissure eruptions of basaltic magmas. In the southernmost part of the Oslo Region these eruptions resulted in a pile of nephelinites, hawaiites, and basanites at least 2000 m thick. Basalt thickness decreases whereas silica-activity increases northwards (and with time) so that in the central Oslo Region this early volcanism is represented by only a single flow of quartz tholeiite, 15-30 in thick. Farther north no early basalts are found. Magmatic activity continued along the entire Oslo Region with extrusions of large volumes of near silica-saturated (trachytic) rhomb porphyry lavas and minor amounts of mildly undersaturated to saturated basaltic lavas and rhyolites from fissures, later from central volcanoes. The magmatic activity ended about 245 Ma ago after emplacement into the upper crust of large intermediate saturated to silicic batholiths, and minor amounts of nepheline syenite (e.g. Ramberg and Larsen 1978; Sundvoll 1978a; Neumann 1980).

All these rocks are enriched in incompatible elements including light rare earth elements (REE). Trace element, $^{87}Sr/^{86}Sr$, and $^{143}Nd/^{144}Nd$ data have been interpreted to show derivation from an "enriched" mantle source region with minor heterogeneities. Some rock types appear to be affected by crustal contamination (e.g. Jacobsen and Wasserburg 1978; Sundvoll 1978a,b).

Felsic rocks of monzonitic composition (larvikites and rhomb porphyries) make up about 40 percent of the exposed surface and subsurface rocks (Barth, 1945). Gravity studies of the Oslo Region (Ramberg 1976) have, however, revealed the existence of large masses of dense rocks at depth along the entire

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Oslo rift. When the estimated volume of these mafic to ultramafic rocks is included in the abundance calculations, the proportion of felsic rocks is reduced to about 10 percent (Ramberg 1976), which, from a mass-balance point of view, is compatible with derivation by fractionation or partial melting of a basaltic parent.

Major and trace element data on the Oslo rift magmatic rocks and their minerals, have revealed the existence of a series of fractionation trends leading from semisaturated intermediate (monzonitic) melts to highly over- and undersaturated paralkaline residuals. This stage of fractional crystallization took place in shallow magma chambers now exposed by erosion. Information about the fractional crystallization leading to the semisaturated monzonitic melts (which are the most mafic rocks found in the large shallow batholiths in the Oslo rift), may be deduced from experimental data. It is suggested that in tectonically quiet periods between periods of high tectonic activity, and during the final stages of the Permian rifting episode in the Oslo Region, basaltic melts rising from the mantle source region were unable to penetrate the less dense crust, and accumulated near the mantle/crust boundary. Here fractional crystallization (controlled by the olivine-pyroxene liquidus boundary in the field of silica saturation) proceeded to the stage at which the residual melts had enough buoyancy to force their way to the surface or the upper crust, thus giving rise to the high proportion of approximately silica-saturated intermediate rocks in the Oslo Region, leaving large masses of dense cumulates together with gabbros in the lower crust.

During the ascent to the upper crust, a thermal divide was established along the diopside-forsterite-anorthite join and the olivine-pyroxene liquidus boundary shifted into the field of silica-saturation. Subsequent low-pressure fractional crystallization forced the evolutionary trends away from the field of silica-saturation, and gave rise to strongly over- and undersaturated derivatives, depending on the silica activity in the melt at the time of ascent. Petrogenetic studies in progress on other rocks in the Oslo Region support this general model.

It is likely that the evolutionary course outlined for the Oslo rift has been followed in many cases of continental rifting, with the pressure in the lower part of a thinned crust playing the dominant role in determining the composition of the evolved rocks. Seismic and gravimetric studies have shown that under active rifts (e.g. the Rio Grande, the East African and the Baikal rifts) a zone of low-velocity, low-density material is typically found below the thinned crust (e.g. Fairhead 1976; Bridwell 1978; Logatchev et al. 1978). Such zones are frequently interpreted as invasion of asthenospheric material up to the mantle-crust boundary. It should be emphasized, however, that such zones of low-velocity, low-density material may equally well represent accumulation near the mantle-crust boundary of melts rising through the solid lithosphere from some source region.

Cooling of melts in the complex system of magma chambers and feeding channels which must have existed in the lower crust during the Permian rifting episode in the Oslo Region would be expected to involve complex processes such as fractional crystallization in periodically refilled magma chambers, partial remelting of previously crystallized gabbros, mixing between magmas, plus contamination and assimilation of lower crust materials. All these processes are likely to obscure geochemical characteristics inherited from the mantle source region(s), and to cause higher degrees of

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enrichment of incompatible elements than does simple fractional crystallization alone. The "enriched" Oslo Region magmatic rocks may thus well be derived from an "undepleted" type mantle source, as suggested by Jacobsen and Wasserburg (1978) on the basis of Sm-Nd and Sr-Rb isotope data.

KEY WORDS: Oslo rift, trachytic rocks, geochemistry, petrogenesis, silica-activity, basalt, Permian, paleorift, fractional crystallization.

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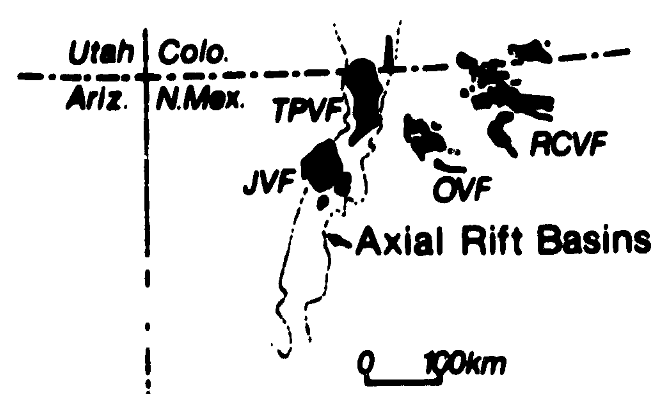
LATE CENOZOIC VOLCANISM OF THE NORTHERN NEW MEXICO-SOUTHERN COLORADO PORTION OF THE RIO GRANDE RIFT. Michael A. Dungan, Dept. of Geological Sci., Southern Methodist Univ., Dallas, TX; David Phelps, Exxon Prod. Res., Houston, TX; John C. Stormer, Dept. Geology, Univ. Georgia, Athens, GA; Douglas P. Blanchard, NASA-JSC, Houston, TX; Roger Nielsen, Nancy McMillian and Ren Thompson, Dept. Geological Sci., Southern Methodist Univ., Dallas, TX.

INTRODUCTION. The volcanic-tectonic evolution of the northern Rio Grande Rift has been highly episodic. The initial period of rifting, which began at 27-25 m.y.b.p., was followed by a period of relative quiessence (17-10 m.y.b.p.). A renewal of volcanism, uplift and differential movement on major basin-bounding faults appears to have begun in the late Miocene. This event may have peaked in mid to late Pliocene time (3 ± 1.5 m.y.b.p.), but it has continued into the late Quaternary. In the northern Rio Grande Rift, volcanism of latest Miocene to Holocene age has been localized along an ENE-trending belt extending from east-central Arizona to northeast New Mexico (Jemez Lineament).

This preliminary report is concerned with the three northeastern-most volcanic fields along this chain. One of the goals of this study is to acquire comparable major and trace element geochemical data on the spectrum of lavas found in each field and compare them as a function of structural position in the rift as it has been broadly defined by Cordell (1978). The Taos Plateau (TPVF), Ocate (OVF) and Raton-Clayton (RCVF) volcanic fields comprise an east-west array extending from the axial graben to the eastern flank of the rift (see figure below). These three fields are broadly time equivalent and show many striking similarities (e.g., Stormer, 1972a). However, as first noted by Lipman (Lipman, 1969; Lipman and Mehnert, 1975), there is a spatial progression from mainly tholeiitic volcanism in the axis of the rift (TPVF) to predominately alkalic volcanism on the northeast flank. New, if fragmentary, data on the OVF suggests that it is chemcially and petrologically intermediate between the other two in keeping with its geographically intermediate position. A similar trend was noted by Lipman in comparing the TPVF basalts to rift age basalts on the western flank (central Colorado-San Juan Mtns.-Hinsdale lavas). Analogous geochemical differences between axial and flank lavas have been recognized from other rifts (e.g., Baker et al., 1972).

Lipman and Mehnert (1975) have suggested that the progressively alkalic character of the flank lavas with increasing distance from the rift axis reflects lateral differences in crust-mantle struture and thermal evolution. These differences arise largely from greater sub-lithospheric heat flow beneath the rift axis resulting in more pronounced asthenospheric uprise, overall uplift and crustal attenuation. We have adopted this hypothesis as a working model which we hope to test by petrologic modeling of a greatly expanded data base.



Late Cenozoic Volcanic Fields Northern New Mexico and Southern Colorado

Raton-Clayton	7.5 - 0.01 m.y.
Ocate	8.1 - 0.8 m.y.
Taos Plateau	4.5 - 1.8

LATE CENOZOIC VOLCANISM OF NORTHERN NEW-MEXICO

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DISCUSSION. The purpose of this note is to update the documentation of some of the temporal and spatial variations in volcanic geochemistry and petrology that exist within and among the TPVF, OVF, and RCVF. Each field is a complex volcanic association including several basalt types and a spectrum of coeval intermediate and silicic rock types. We have restricted the discussion to mafic and intermediate rock types as the silicic volcanics (>65% SiO_2) are minor and difficult to compare among fields.

Utilizing our as yet to be completed data base, we have tabulated several chemical parameters in Table 1 and divided the rocks into five groups. This grouping involves some arbitrary boundaries between rock types that appear gradational on the basis of our data, and ignores a few single samples which fail to fall within the normal limits for the other rock types. Group IV in particular contains a wide diversity of rocks which we will eventually subdivide. The characterization of OVF samples lags well behind that of the other two fields, suggesting possible revisions.

Group	Field	Local Rock Name	SiO_2	Q/Ne	Hy	Mg#	K_2O	Na/K	La/Sm	La_N	Ba	Sr	P_2O_5	TiO_2
I	RCVF	Mafic Feldspathoidal	36-45	8-30	-	66-69	0.9-1.7	4-3	7-10.5	300-825	1000-2200	1500-3500	1.0-3.0	1.3-2.0
II	OVF	Alkali Olivine Basalt	46-49	<7	-	65-60	1.5-2.0	2.5-3.5	6.5-7.5	125-150	550-1400	1000-1500	0.5-0.6	1.3-1.7
	RCVF	Alkali Olivine Basalt	46-47	<10	-	60-50	1.1-1.6	1.8-2.5	6.2-8.5	160-400	1000-2100	950-2500	0.8-1.2	1.7-2.0
III	TPVF	Servilleta Tholeiites	49-52	-	5-20	59-55	<1.0	6-3	2.5-4.0	20-55	200-500	280-450	0.15-0.25	1.1-1.3
	OVF	Low K Basalt	49-53	?	<10	60-55	<1.1	6-3	4.0-5.5	65-90	350-1200	500-800	0.2-0.3	1.3-2.0
	RCVF	Low K Basalt	48-51	?	<10	65-55	<1.0	6-3	3.7-7.1	50-100	360-800	450-1100	0.25-0.80	1.1-1.7
IV	TPVF	Silicic Alkalic Basalt	50-52	-	<12	60-55	1.2-2.0	1.8-2.5	4.3-5.0	60-140	550-2000	600-1200	0.1-0.9	1.2-1.8
	TPVF	Xenoc. Basaltic Andesites	55-58	<5	10-20	55-45	1.6-2.3	1.5-1.9	4.8-5.2	90-170	1000-1400	600-800	0.5-0.7	1.3-1.4
	OVF	Silicic Alkalic Basalts	50-52	-	<12	60-45	1.5-2.0	1.5-2.5	5.3-7.0	100-120	700-1200	600-1100	0.2-0.4	1.5-1.8
	OVF	Basaltic Andesites	53-56	<3	5-17	60-45	1.6-2.4	1.5-2.5	5.5-7.5	110-130	700-2500	450-750	-	1.3-1.8
	RCVF	Capulin lavas	51-56	<5	5-17	60-52	1.2-2.4	2-3	5-8	75-170	550-1400	450-850	0.3-0.6	1.2-1.5
V	TPVF	Basaltic Andesite	54-57	<5	>10	57	2.3	2.5	6	150	1400	1100	0.35	1.25
	TPVF	Olivine Andesite	57-60	3-10	>13	59-50	2.1-2.8	2.0	6-7	120-160	1000-1600	800-1300	0.4-1.0	0.7-1.3
	TPVF	Dacite	62-64	10-20	<5	52-43	2.7-3.3	1.5	6.5-7.5	110-150	1100-1400	500-1600	0.1-0.7	0.7-0.9
	RCVF	Sierra Grande Andesites	57-60	10-12	5	53-56	2.4-2.8	2	9.5	185	1500-1700	1000-1400	0.5-0.6	0.8-1.0

GEOLOGY, GEOCHRONOLOGY AND TEMPORAL VARIATIONS. The TPVF occurs within the San Luis grabed. Early rift-age (22-25 m.y.) volcanics are remnants of andesite to rhyolite stratovolcanoes. With the exception of a 10 m.y. quartz latite dome, all the exposed latest Cenozoic volcanics are 4.5-1.8 m.y. (Lipman and Mehnert, 1979). Rio Grande Gorge exposures demonstrate that Group III, IV and V lavas were erupted within the same time span (4-2.8 m.y.). The Servilleta olivine tholeiites (III), olivine andesites and dacites (V) all form large monolithologic shields. The Sierra Grande andesite (V-RCVF) also occurs as a large shield. Group IV lavas are almost entirely restricted to single cinder cone-flow eruptions.

The OVF occupies the eastern flank of the Sangre de Cristo Mountains whereas the RCVF occurs on the high plains to the east. In both fields, extensive lava flows cap old erosional levels. The oldest flows occupy the highest terrace surfaces (>5 m.y.) and progressively younger flows occur at successively lower elevations. The presence of 5.8-8.5 m.y. flows at 10,000 ft. in the Sangre de Cristos attests to the concurrence of volcanism and tectonism during the latest Cenozoic. In both the OVF and RCVF the volcanism appears to have occurred in discrete pulses:

OVF-- 8.5-5.8 m.y., 4.8-4.0 m.y., 3.3-3.0 m.y., 2.2 m.y., 1.4 m.y. and 0.8 m.y. (O'Neill and Mehnert, 1980)

RCVF-- 7.2±.5 m.y., 3.5±.5 m.y., 2.5±.3 m.y., 1.8-1.9 m.y. and 0.01 m.y. (Stormer, 1972b).

The latest Miocene-early Pliocene volcanics in the RCVF are almost entirely comprised of alkali olivine basalts (II). Rocks of the same age in the OVF are more diverse; the early phase of volcanism includes representatives of Groups II and III, but is dominated by silicic alkalic basalts (IV). No rocks of this age are exposed in the TPVF. However, they may very well be buried beneath valley fill.

The maximum diversity of volcanic rock types is developed in all three fields suring the main phase of Pliocene volcanism (4.5-1.8 m.y.b.p.). This is the only phase represented in the TPVF and accounts for >50% of the volcanics in the OCF and RCVF. Group III basalts are the dominant mafic magma type during this time period, particularly in the TPVF. Shield volcanoes of Group V lavas are entirely restricted to the latest Pliocene ~3.5-1.8 m.y.

The last, waning stages of volcanism in all three fields are dominantly or entirely comprised of Group IV lavas. These consist of cindercone-flow eruptions of silicic alkalic basalts and xenocrystic basaltic andesites, and in the case of the RCVF very minor amounts of nephelinite. Available age dates indicate that volcanism continued on the east flank of this portion of the Rio Grande Rift after it ceased in the axial graben (TPVF 1.8 m.y.; OVF 0.8 m.y.; RCVF .01 m.y.). The Brazos volcanics, whicn occur 20-30 km west of the TPVF are also young (0.25 m.y.) Group IV lavas.

SPATIAL VARIATIONS IN VOLCANIC GEOCHEMISTRY. This section is a comparison of the rock types among and within the volcanic fields based on three criteria: (1) restriction of certain rock types to specific volcanic fields; (2) volume relations; and (3) geochemical variations within groups as a function of distance from the rift axis.

Groups I and II: Basanites and nephelinites are present only in the RCVF and ne-normative rocks are absent from the TPVF. There are minor alkali olivine basalts in the OVF, but these are much more abundant in the RCVF. Geochemical data from the OVF are insufficient to support a detailed comparison with the RCVF, but available data suggest higher concentrations of LIL elements in Group II basalts in the RCVF.

Group III: These rocks are well represented in all three fields, but are the volumentrically dominant rock type only in the TPVF (~80%). The Servilleta olivine tholeiites have the lowest concentrations of LIL elements and lowest La/Sm of any rock type recognized. Concentrations of LIL elements and La/Sm increase to the east in the low K basalts of OVF and RCVF. The intra-group geochemical variations within Group III reflect an apparent tendency towards increasingly alkalic affinities from rift axis to flank. This group embodies the most consistent and well-defined west to east geochemical trends of any troup.

Group IV: Basalts and basaltic andesites (typically xenocrystic) of transitional affinity are the volumetrically dominant rock types only in the OVF. In the TPVF these rocks are far subordinate in volume to III and V lavas and are a minor percentage of the highly diverse RCVF. Only in the latest stages of volcanism are these lavas more abundant than other types. In keeping with their transitional normative character, these lavas are generally intermediate between Group III and II lavas in their LIL element concentrations. The Group IV lavas show a trend of increasing LIL abundances from rift axis to rift flank which is similar to, but not as clearly defined as the trend shown by the Group III basalts.

Group V: This group is abundantly represented only in the TPVF where olivine andesites and two pyroxene dacites comprise 12 major shields. Group V rocks have not been identified in the OVF. The Sierra Grande ahdesite (2 pyroxenes) is the lone representative in the RCVF. Group V andesites overlap in terms of LIL element concentrations with Groups II and IV but are quite distinct in major elements (esp. SiO_2). The origin of the TPVF olivine andesites is a major enigma as they appear to require a high-SiO_2 parent magma or an unusual fractionation scheme for which there is no physical evidence. The dacites appear to be generated by mixing of olivine andesites and a lower crustal granulite component (Williams and Murthy, 1979).

CONCLUSIONS. Important variables in determining the compositions of basaltic parent magmas are: (1) depth of melting; (2) degree of partial melting, and (3) composition of the mantle. All three of these appear to have been important in generating the spectrum of observed magma types. The variations in basalt geochemistry as a function of position relative to the rift axis are presumed to be primarily due to differences in the depth and degree of partial melting. These in turn reflect lateral thinning of the lithosphere approaching the rift axis. If it may be assumed that mantle diapirs rise more rapidly and to shallower levels near the rift axis, the general aspects of basalts in the various fields may be qualitatively explained.

The dominance of tholeiitic basalts near the rift axis is consistent with segregation of basalts at relatively shallow levels compared to the rift flank where there is an increasing abundance of understaurated magmas. However, several source regions must have been tapped in each field in order to account for the more or less synchronous eruption of three or more mafic magma types. The association of melilite nephelinites with rocks of tholeiitic affinity (III) in the RCVF suggests that the source regions were separated by a large depth interval. The successive elimination of highly then moderately undersaturated rocks as the rift axis is approached, is suggestive of a decreasing depth range of source regions and a shallower maximum depth of magma segregation. The earlier cessation of volcanism in the TPVF may reflect a more rapdily rising diapir which produced more efficient magma segregation at the rift axis.

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Late Cenozoic Basalts of the Southern Rio Grande Rift, Southern New Mexico, West Texas, and Northern Chihuahua, Mexico

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Introduction

Numerous outcrops of alkali basalt occur in southern New Mexico, west Texas and northern Chihuahua; the majority of the basalts occur in the rift zone west of the present Rio Grande (Fig. 1). Seager and Morgan (1979) state that extrusive basalt appeared about 13 million years ago in the area and that about 2-3 million years ago the vulcanism accelerated possibly because of thinning of the crust beneath the rift. Seager and Morgan (1979, in Fig. 2) place the western boundary of the rift just east of Deming, New Mexico, which would place the basalts near Columbus, Palomas arroyo, Hachita, and Deming west of the rift.

Lipman (1969) first pointed out that basalts of different chemistry occur within and on the flanks of the northern Rio Grande rift. Basalts in the northern rift are typically tholeiite with lower total alkalis, titanium, and phosphorous compared to the more alkali basalt west and east of the rift.

In the southern part of the rift basalt studies have been in progress for the past 12 years (Hoffer, 1971, 1972, 1975, 1976, 1981, Frantes, 1981, and Sheffield, 1981). Although not all the basalt occurrences have been studied, 7 of the 11 occurrences shown in Fig. 1 have been completed or are currently under investigation.

Basalt Chemistry

The basalts of the southern Rio Grande rift are alkali which, except for the Palomas volcanics, show moderate to little differentiation. Basalts that occur within the rift display significantly low concentrations of SiO_2 (45.6% versus 49.1%), higher TiO_2 (2.3% versus 1.7%), and lower NaO_2 (2.9% versus 3.6%). The higher SiO_2 and Na_2O values may be the result of contamination of the basalt magma erupted through the thick, little faulted continental crust on the flanks of the Rio Grande rift (Lipman, 1969).

The basalts showing the highest degree of differentiation are the Palomas volcanics, on the west flank of the rift. The volcanic units include older picrites and olivine basalts overlain by hawaiites, trachybasalts, mugearites, and tristanites (Frantes, 1981). Compositionally, total alkali content of the rocks range from 3.6% to 8.4%, silica from 41.8% to 63.6% and magnesium oxide from 1.2% to 19.4%.

Basalts of the Southern Rio Grande Rift

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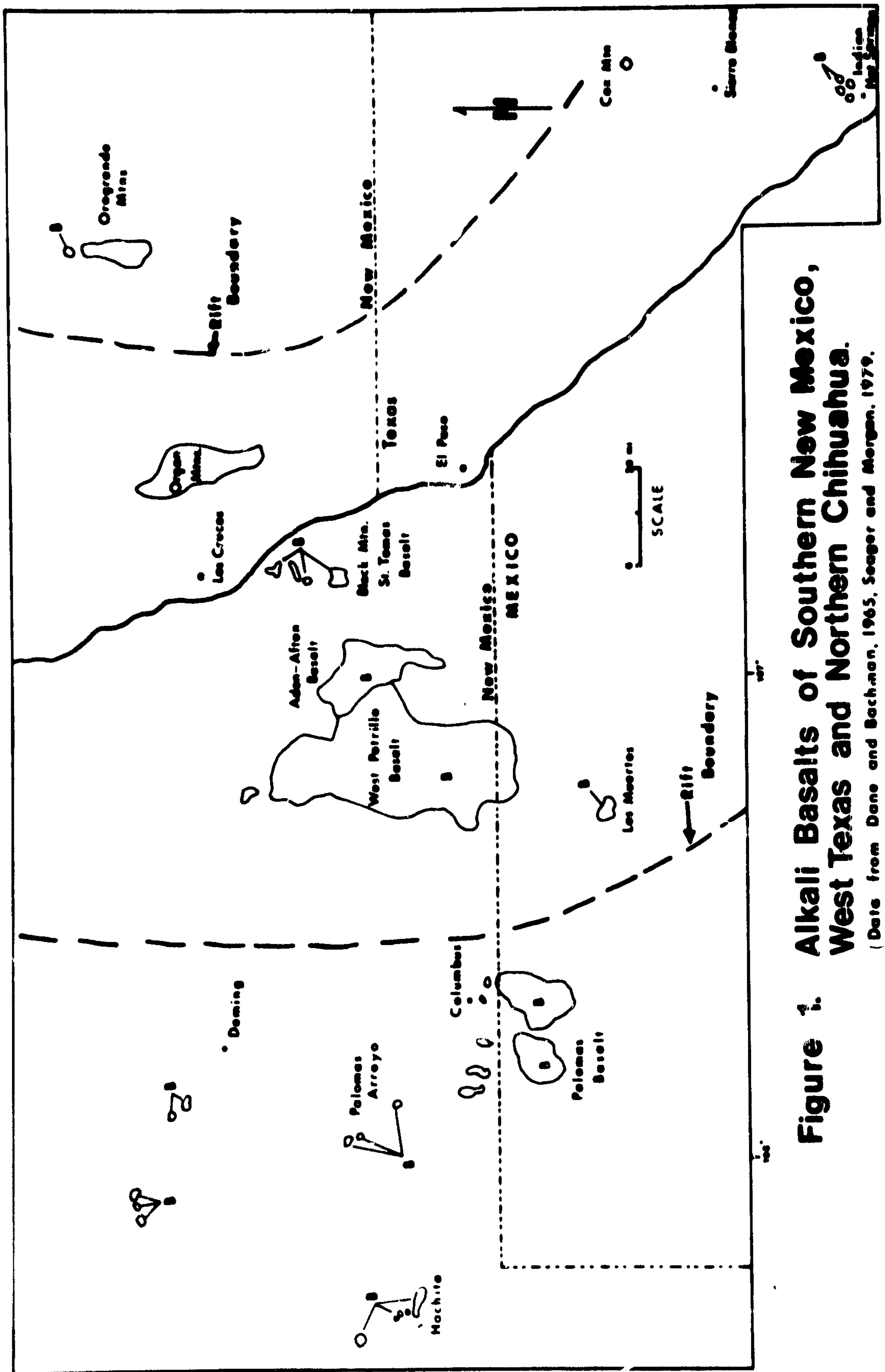

Figure 1. Alkali Basalts of Southern New Mexico, West Texas and Northern Chihuahua.
(Data from Dane and Bachman, 1965, Seager and Morgan, 1979, Albritton and Smith, 1965, Jones and Reeser, 1970. B-basalt

Basalts of the Southern Rio Grande Rift

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Basalt Ages

Age dates of the basalts have been obtained from only five areas (Table 1). These dates show that the basalts are progressively older toward the west from the Black Mountain-Santo Tomas chain to the Palomas volcanic field.

Seager and Morgan (1979) state that rifting initiated about 28 m.y. ago with crustal extension producing a thinning of the crust or lithosphere accompanied by a mantle bulge which initiated the resulting basalt volcanism. Based upon the increasing age dates of the basalts westward across the rift it is suggested that crustal extension was accompanied by eastward migration of a mantle plume.

Location	Bk Mtn.- S. Tomas Rift 1-5 mi West	Aden-Afton Rift 20 mi West	West Potr. Rift 30 mi West		Palomas Out of Rift 60 mi West	Hachita Out of Rift 105 mi West	Deming Out of Rift 65 mi West	Sierra Blanca Out of Rift 5 mi East
Age (millions of years)	0.23 0.55	0.18- 1.23	1.23[2]		2.96[2] 5.17[2]	n.d.	n.d.	16.3[1]
Chemistry (samples)	(15)	(12)	(75)		(57)	(4)	(2)	(1)
			M1	M2				
SiO_2	47.0	45.2	45.2	44.9	51.6	48.8	48.0	48.0
TiO_2	2.4	2.0	2.5	2.4	1.8	1.7	1.5	1.9
Al_2O_3	15.9	14.6	14.5	14.3	15.0	14.8	15.7	15.2
FeO_3[3]	11.7	12.3	12.5	11.9	9.5	12.6	11.5	11.17
MnO	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2
MgO	7.7	9.8	8.7	10.4	5.9	8.5	8.3	8.2
CaO	10.0	10.5	10.7	10.0	7.2	9.6	10.4	8.2
Na_2O	2.9	3.0	2.9	2.8	4.2	3.4	3.3	3.3
K_2O	1.8	1.5	1.4	1.4	2.1	1.7	0.7	1.7
P_2O_5	0.6	0.5	0.7	0.7	0.6	0.5	0.5	0.7
S.I. 4	32	37	34	39	5-57	32	34	38

Table 1. Characteristics of basalts from the southern Rio Grande Rift (1-Barker, 1980; 2-Hawley, 1981; 3-total Fe reported as FeO; 4-S.I. = Solidification Index, M1 = Member 1, M2 = Member 2, n.d. = not dated).

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THERMAL AND RHEOLOGIC HISTORY OF THE UPPER MANTLE BENEATH THE SOUTHERN RIO GRANDE RIFT: EVIDENCE FROM KILBOURNE HOLE XENOLITHS

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Alkalic basalts containing mantle xenoliths have been erupted in many young continental rifts. At Kilbourne Hole, N.M. xenoliths of both crustal and mantle rocks are especially abundant, and allow important constraints to be placed on processes operative beneath the southern Rio Grande rift. We have studied both the mineral chemistries and textures of a suite of spinel lherzolites in the belief that a realistic understanding of upper mantle evolution requires consideration of both chemical and rheologic processes.

Two textural types of spinel lherzolite are present at Kilbourne Hole: fine grained, tabular equigranular and coarse grained protogranular. Both of these types (especially the former) are cut by spinel pyroxenite dikes (as seen in composite xenoliths), but no structural relationship between the two lherzolite types has been observed. The two types differ in bulk chemical composition, the tabular equigranular samples being characterized by generally higher Fe and light REE contents (Irving, 1980).

Petrofabrics

Olivine lattice orientations of the tabular equigranular xenoliths have a strong orthorhombic symmetry, whereas those of the protogranular xenoliths show a pronounced uniaxial symmetry about [010] - see Fig. 1. Experimental studies in which both textures and corresponding olivine lattice orientations have been reproduced indicate a diversity of deformation, recrystallization and grain growth characteristics for these samples. The well developed [010] axial symmetry for the protogranular xenoliths is comparable to the fabric pattern obtained by Carter and Avé Lallemant (1970) in natural dunites and lherzolites deformed between 1100-1200°C and 13-15 kb at strains below 20% in the presence of externally released water. This suggests that the chemically depleted protogranular xenoliths have undergone syntectonic recrystallization in the presence of intercrystalline fluid (possibly a partial melt and/or a metasomatic fluid). The strong orthorhombic symmetry for the tabular equigranular xenoliths is similar to that of a harzburgite from the ultramafic section of the Samail ophiolite (Boudier and Coleman, 1981). The olivine lattice orientations for the tabular equigranular wall rock of a composite xenolith have a dominant orthorhombic pattern, however weak girdles about a weak [010] maximum are also discernible and may result from the formation of the pyroxenite dike.

Geothermometry

Olivine-spinel (Fabriès, 1979) and two-pyroxene (Wells, 1977) thermometers show agreement for the protogranular xenoliths and yield apparent equilibration temperatures between 950°C and 1050°C. Conversely, large discrepancies between the two methods are found for the tabular equigranular xenoliths (see Fig. 2). In these samples the spinel is demonstrably out of equilibrium with the silicate phases, however neither this disequilibrium nor inherent differences between the thermometers can account for such large discrepancies. Rather we suggest that a recent thermal perturbation (up to 1150°C) has affected the tabular equigranular samples, and this is further supported by Ca zoning profiles in olivine. Possibly this event is linked to the intrusion of the host basanite prior to and during xenolith entrainment and ascent.

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THERMAL/RHEOLOGIC HISTORY OF UPPER MANTLE

Bussod, G.Y. and Irving, A.J.

Geobarometry

The absence of both plagioclase and garnet constrains the source region of the spinel lherzolites to pressures between 8 and 24 kb. Finer resolution within the spinel lherzolite facies is possible by consideration of solubility of Al_2O_3 in orthopyroxene (Dixon and Presnall, 1980) and of Ca in olivine (Finnerty and Boyd, 1978). Pressures calculated using both barometers are plotted against each other in Fig. 3. The lack of any linear relationship between the results attests to the qualitative nature of these estimates. Nevertheless, there is a reasonably clear distinction between the protogranular and tabular equigranular types which suggests that the latter are stratigraphically lower and that the upper mantle beneath Kilbourne Hole is texturally and chemically stratified.

Discussion

Our P-T estimates using olivine and pyroxenes define a steep minimum mantle geothermal gradient for the southern Rio Grande rift (see Fig. 4). Similar estimates were obtained by Reid (1976). A lower crustal geotherm determined by Padovani and Carter (1977) using granulite xenoliths from Kilbourne Hole is in agreement with a best fit geotherm of about 30°C/km calculated by Bridwell (1978). These gradients are applicable to the time when the crustal xenoliths were brought to the surface (5.0 to 0.1 my ago), yet the crustal gradient is remarkably consistent with the value predicted from surface heat flux measurements (Decker and Smithson, 1975). However, it is very difficult to connect the crustal and upper mantle geotherms (see Fig. 4). Seager and Morgan (1979) have suggested that two thermal regimes may exist: one in the lower crust, perhaps related to large-scale intrusion of basaltic dikes, and one in the upper mantle, perhaps related to a convective mantle diapir.

The Kilbourne Hole xenoliths provide direct evidence that the upper mantle beneath this region is chemically and rheologically zoned. Composite xenoliths interpreted as wall rock and precipitates from intrusive magma reinforce the concept of an intrusive complex at depth. The most complex thermal histories are preserved in the shallower, tabular equigranular xenoliths, which are chemically enriched in a basaltic component. This may be explained by the presence of an intrusive basaltic system in the lower crust and uppermost mantle composed of dikes and sills where advective transfer of heat to the crust and immediately underlying mantle is responsible for local chemical and rheologic changes through time. It is postulated that a brittle-plastic transition regime exists near the crust-mantle boundary such that at progressively shallower depths more intrusive bodies are stable along the rift.

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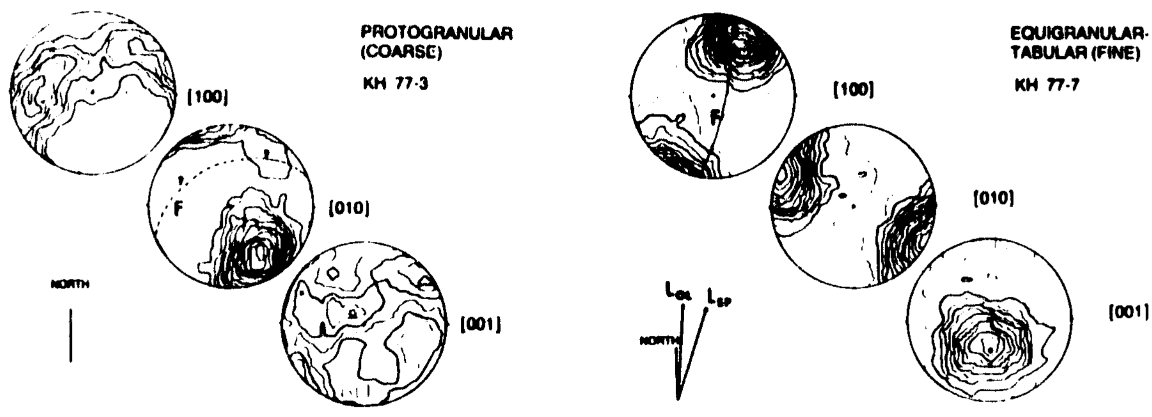

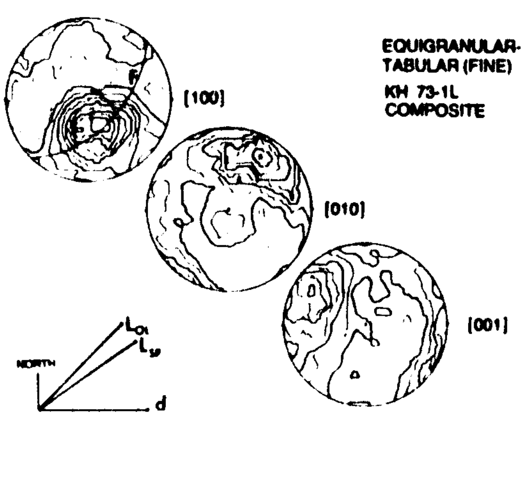

Fig.1. Kambplot orientation density diagrams for olivine grains in xenoliths

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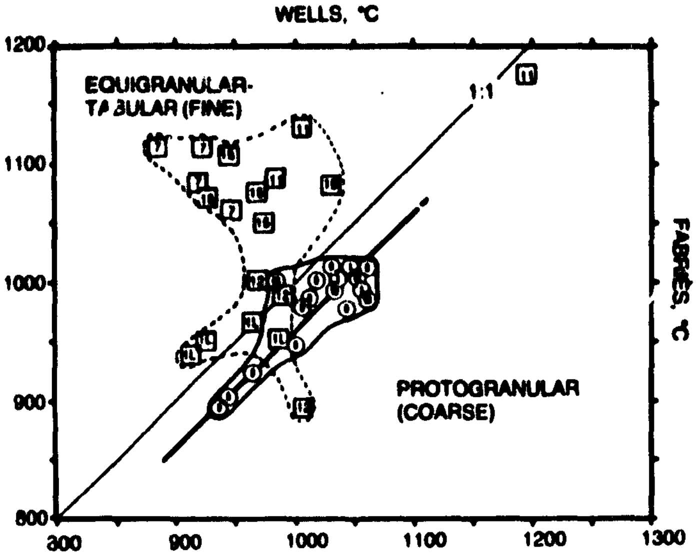

Fig. 2. Temperature estimates

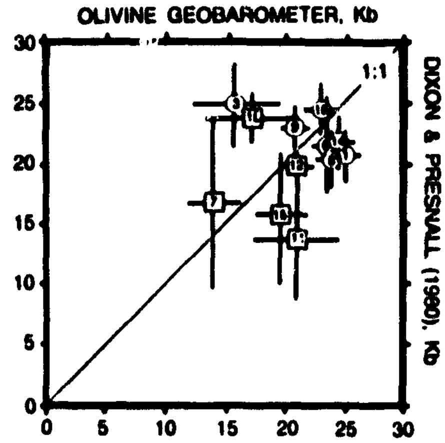

Fig. 3. Pressure estimates

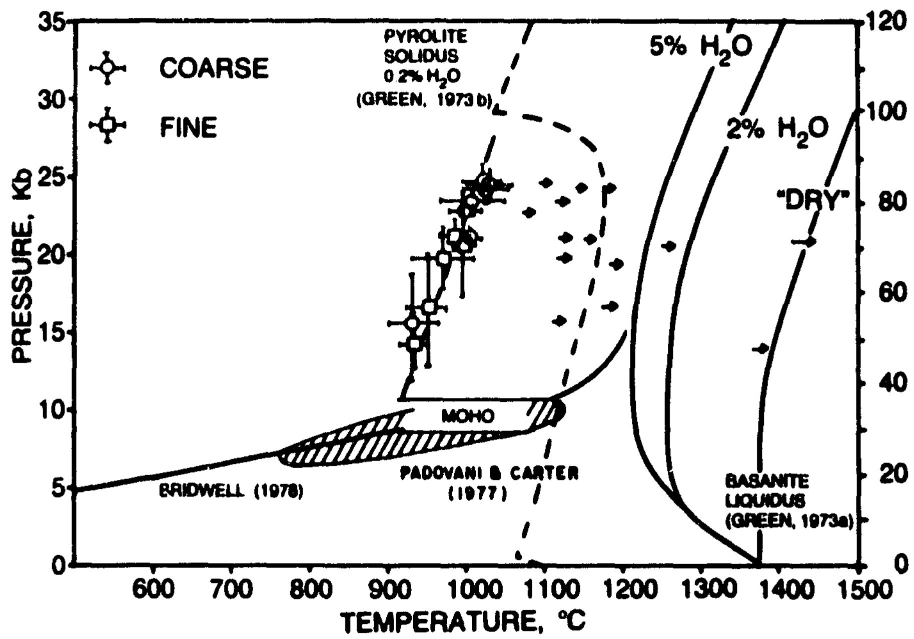

Fig. 4. Minimum geothermal gradient for Kilbourne Hole xenoliths. Arrows represent temperature estimates of gradient during xenolith entrainment based on calcium contents of olivine rims (assuming constant pressure)

GEOCHEMICAL CONSTRAINTS ON THE EVOLUTION OF THE LOWER CRUST BENEATH THE RIO GRANDE RIFT, E.R. Padovani and S. R. Hart, Dept. Earth & Planetary Sciences, Massachusetts Inst. of Technology, Cambridge, MA 02139

As part of a long-range objective to characterize the lower continental crust, petrologic and geochemical studies have been undertaken on representative suites of lower crustal xenolith samples from Kilbourne Hole maar and Potrillo maar in south central New Mexico. Major element analyses completed on about 50 samples indicate a wide range of compositions among the orthogneisses and paragneisses, with SiO_2 contents ranging from 43% to 70%, Na_2O contents ranging from less than 0.5% to 6% and K_2O contents ranging from 0.2% to 6%. The data do not suggest any dramatic depletion of alkalies in these granulite facies rocks. Rb/Sr and Sm/Nd systematics reveal mineralogic and isotopic disequilibria over a 1-3 cm scale defined by compositional layering within xenoliths. Although disequilibrium exists between layers, mineral pairs (such as plagioclase-garnet, plagioclase-clinopyroxene and K feldspar-plagioclase) are commonly in perfect equilibrium which must have occurred within the past 10-20 m.y. Monotonic Sr isotopic gradients within paragneisses may hold the key to understanding a much larger scale diffusion gradient (larger than individual xenoliths) at depth which may explain the difference in behavior between the Rb/Sr and Sm/Nd systems. The "errorchron" defined by the xenolith suite is consistent with an age of 1.7 b.y. (see Fig. 1) for the basement beneath the southern Rio Grande Rift. The combination of mineral geothermometers-geobarometers and isotopic results allows construction of a tentative time-temperature history for the lower crust beneath the rift and distance-scales of isotopic exchange.

Time-temperature history. The original age of cratonization in this area was about 1.7 b.y. Granulite conditions probably prevailed in the lower crust during this orogenic event. Following orogenesis, the crust cooled to perhaps stable shield geothermal gradients (implying lower crust temperatures of less than or equal to 500°C). Based on diffusion models, we believe the inter-layer isotopic gradients would not survive continuous granulite conditions for the whole period 1.7 b.y. to present. Other intrusive events have been recognized in nearby regions of New Mexico, Texas and Mexico, with ages of 1.6, 1.2-1.7 and 0.5 b.y. (Muehlberger and Denison, 1964; Silver et al., 1977; Loring and Armstrong, 1980)--their effect on Kilbourne Hole crust is yet unknown. No significant vertical motion (and erosion) has effected Kilbourne Hole crust during the Precambrian as suggested by the presence (Franklin Mts.) of a shallow granite-rhyolite terrain, perhaps 1 b.y. old. During the cooler, orogenically quiescent Precambrian times, mineral geothermometers would be reset to sub-granulite temperatures. Starting about 30 m.y. ago, with the beginning of extension of the Rio Grande Rift, magmatic activity and crustal thinning increased temperatures again into the granulite range. Peak temperatures of 1000-1100°C were reached, as evidenced by 2-pyroxene thermometry, and "bulk" K-feldspar-plagioclase thermometry. The present granulite textures were probably formed at this stage, along with mobilization of Rb and some Nd isotopic exchange between layers. This peak temperature was followed by some cooling as rifting slowed (about 5 m.y. ago), to bring the geotherm to its present value (Lachenbruch and Sass, 1977; Cook et al., 1978); heat flow models predict present lower crustal temperatures of 800-900°C. K-feldspar "host"-plagioclase thermometry consistently records temperatures of 800-900°C, in agreement with the heat flow models. These thermal conditions were sufficient to keep adjacent minerals in continuing local isotopic equilibrium for Sr and Nd (see Fig. 2). Prior to eruption, probably by heating of the erupting magma, the xenolithic material was brought to >1000°C, such that subsequent decompression during eruption produced local "decompression melt-ting", especially of garnet (Padovani and Carter, 1977b). This heating did not have any significant effect on the isotopic systems, or on mineral chemistry (as witnessed by only small 20-30μ diffusion gradients in plagioclase adjacent to the decompression melt, and the very heterogeneous composition of the melts).

Chemistry of the lower crust. At least under Kilbourne Hole, the lower crust has not been massively depleted in alkalies, and paragneisses retain typically sedimentary isotopic signatures for Sr, Nd and $\delta^{18}O$ (James et al., 1980). Compositional heterogeneity is marked, with a whole spectrum of rock types ranging from very basic orthogneisses to highly siliceous peraluminous paragneisses.

Distance-scales of isotopic exchange. The recent granulite-facies event under Kilbourne Hole, though admittedly not particularly long-lived (≤30 m.y.), has been insufficient to bring about Sr and Nd isotopic homogenization on anything but a rather local (0.1-1 cm) scale. Further thinking is needed to establish the relevance of this result to the question of isotopic heterogeneities in the mantle source-regions of basalts (see Hofmann and Hart, 1978).

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GEOCH. CONST. EVOL. LOW. CRUST BENEATH RIO GRANDE RIFT

Padovani, E. R. and Hart, S. R.

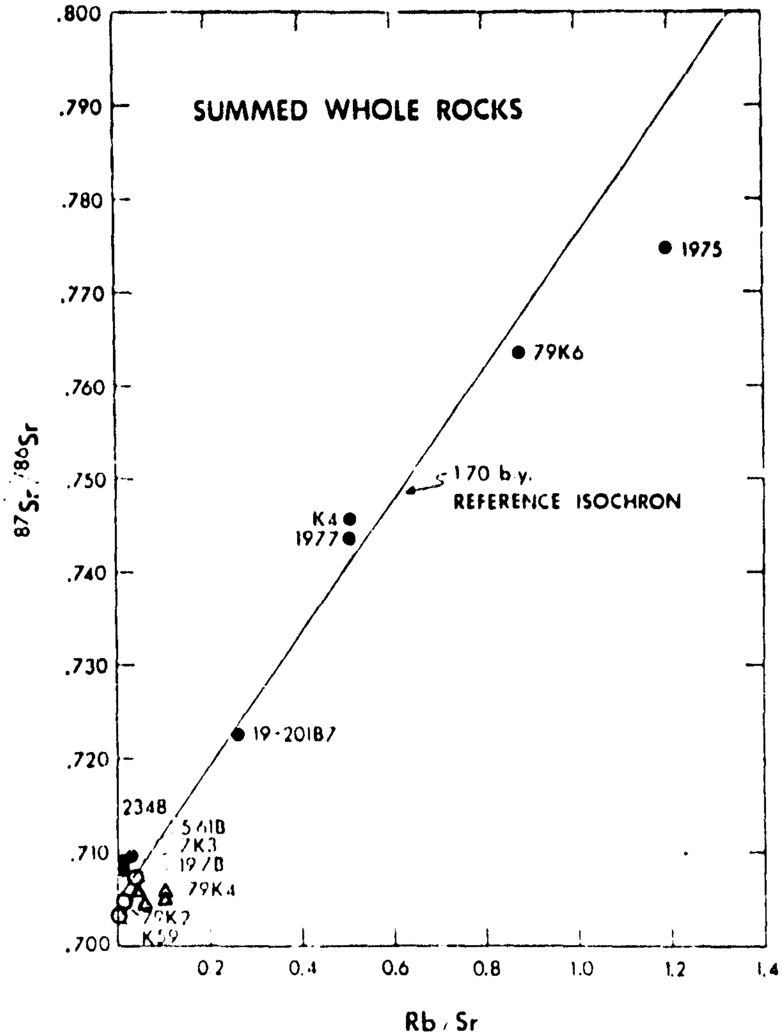

Figure 1. Rb/Sr systematics of "whole rock" xenoliths, obtained either by whole rock analysis of single xenoliths or by "summing" the layers of those xenoliths for which multi-layer Rb/Sr studies were performed (e.g. Fig. 2). Only xenolith 1975 falls significantly away from the 1.7 b.y. reference isochron. This data shows that the "original" age of cratonization of Kilbourne Hole lower crust was ~1.7 b.y. (though this point clearly needs further documentation). Furthermore, though the inter-layer data (such as in Fig. 2) suggests significant relative mobility of Rb versus Sr, the overall effect is not consistent with any dramatic Rb depletion for the lower crust, and the paragneiss samples (closed circles) in particular still show relatively high Rb/Sr ratios. The orthogneiss xenoliths (open circles) are characterized by uniformly low Rb/Sr ratios, but this is probably an original characteristic inherited from their igneous protolith, as opposed to an effect due to large-scale Rb depletion during granulite metamorphism. Note the data for five amphibolite-facies xenoliths from nearby Potrillo Maar (triangles).

GEOCH. CONST. EVOL. LOW. CRUST. BENEATH RIO GRANDE RIFT

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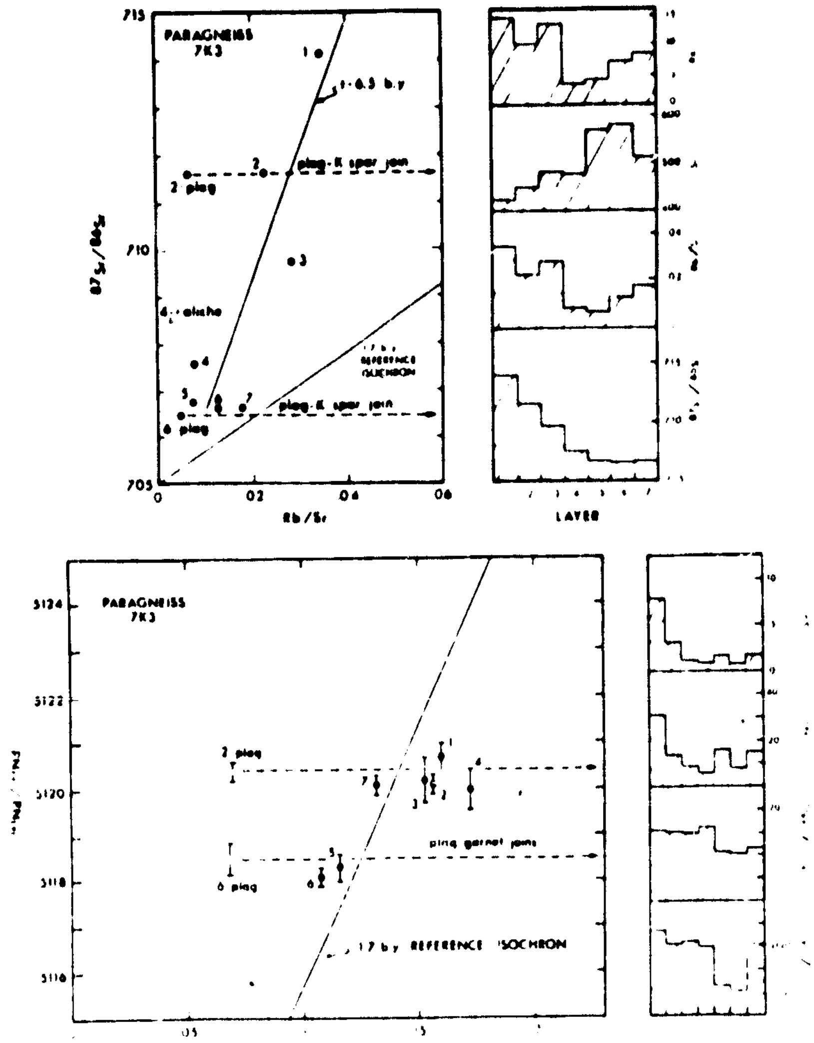


Figure 2. Rb/Sr (top figure) and Sm/Nd (bottom figure) isotopic relationships in minerals and layers of paragneiss 7K3. Minerals from within a given layer are essentially perfectly equilibrated with respect to Sr (plagioclase-K-feldspar pairs from layers 2 and 6) and Nd (plagioclase-garnet pairs from layers 2 and 6); allowing for the observed analytical errors, the maximum ages indicated for these mineral pairs are in the range 10-20 m.y. for both the Sr and Nd systems. In contrast to this small-scale intra-layer isotopic homogeneity, both Sr and Nd show significant isotopic disequilibria between layers. However, with respect to the presumed original age of metamorphism of this paragneiss (~1.7 b.y.), both the Sr and Nd systematics appear to show open-system behavior. The location of layers 1-4 to the left of the reference Sr isochron suggest loss of up to 80% of Rb from these layers at a relatively recent time; this Rb loss has produced a crude linear array of data points with an obviously meaningless slope age of ~6.5 b.y. In contrast, the Nd data suggests some inter-layer isotopic equilibration, as layers 5 and 6 and layers 1-4 and 7 are essentially in isotopic equilibrium. Note the regular monotonic change in $^{87}Sr/^{86}Sr$ across the layers; this is a feature which has been noted in all of the paragneiss samples studied thus far.

CRUSTAL MAGNETIZATION BENEATH THE RIO GRANDE RIFT BASED ON XENOLITHS FROM KILBOURNE HOLE AND POTRILLO MAAR; P. J. Wasilewski, NASA/Goddard Space Flight Center, Laboratory for Extraterrestrial Physics, Greenbelt, Maryland 20771; E. R. Padovani, Dept. Earth and Planetary Sciences, Massachusetts Institute of Technology, Cambridge, MA 02139

Results of magnetic studies on xenoliths from the Colorado plateau and Rio Grande Rift support the concept of the continental Moho as a magnetic boundary (Wasilewski et al., 1979; Wasilewski and Padovani, 1981). Upper mantle peridotites contain chromites which are nonmagnetic at Moho depths. The xenolith results indicate that regional magnetic anomalies in the crust are related to both the topology of the Curie isotherm and to petrologic variations. These results combined with detailed petrologic and geochemical studies on lower crustal xenoliths have revealed that in an area of steep geothermal gradient, the magnetic bottom is at considerably more shallow depths (10-15 km) than is the case in an area with moderate geothermal gradient such as the Colorado plateau where the magnetic bottom is deeper (30-40 km). This is due to the steeper geothermal gradient as well as a different magnetic mineralogy. Though the rift is of limited areal extent it can be recognized in both POGO and Magsat magnetic anomaly maps due to the contrast with surrounding regions.

Beneath the southern Rio Grande rift, the crust appears to be more reducing with increasing depth as reflected by the ilmenite dominated anhydrous lower crustal xenoliths at Kilbourne Hole which have Curie points less than 300°C and characteristically small saturation magnetization and remanence. In contrast some of the xenoliths from Potrillo maar which are considered to represent intermediate crustal depths between those defined by exposed outcrop and wells drilled to basement and those defined by the granulite facies xenoliths have 550°C Curie points and large values of saturation magnetization and remanence. Granulite xenoliths from Elephant Butte and the Lucero Volcanic field (Wasilewski and Baldrich - unpublished research) have magnetic characteristics that differ from both the Colorado plateau and Kilbourne Hole and Potrillo maar granulites suggesting different conditions of equilibration may exist at lower crustal depths along the rift. It appears that active regions with high heat flow such as rifts may be anomalous with respect to their magnetic properties.

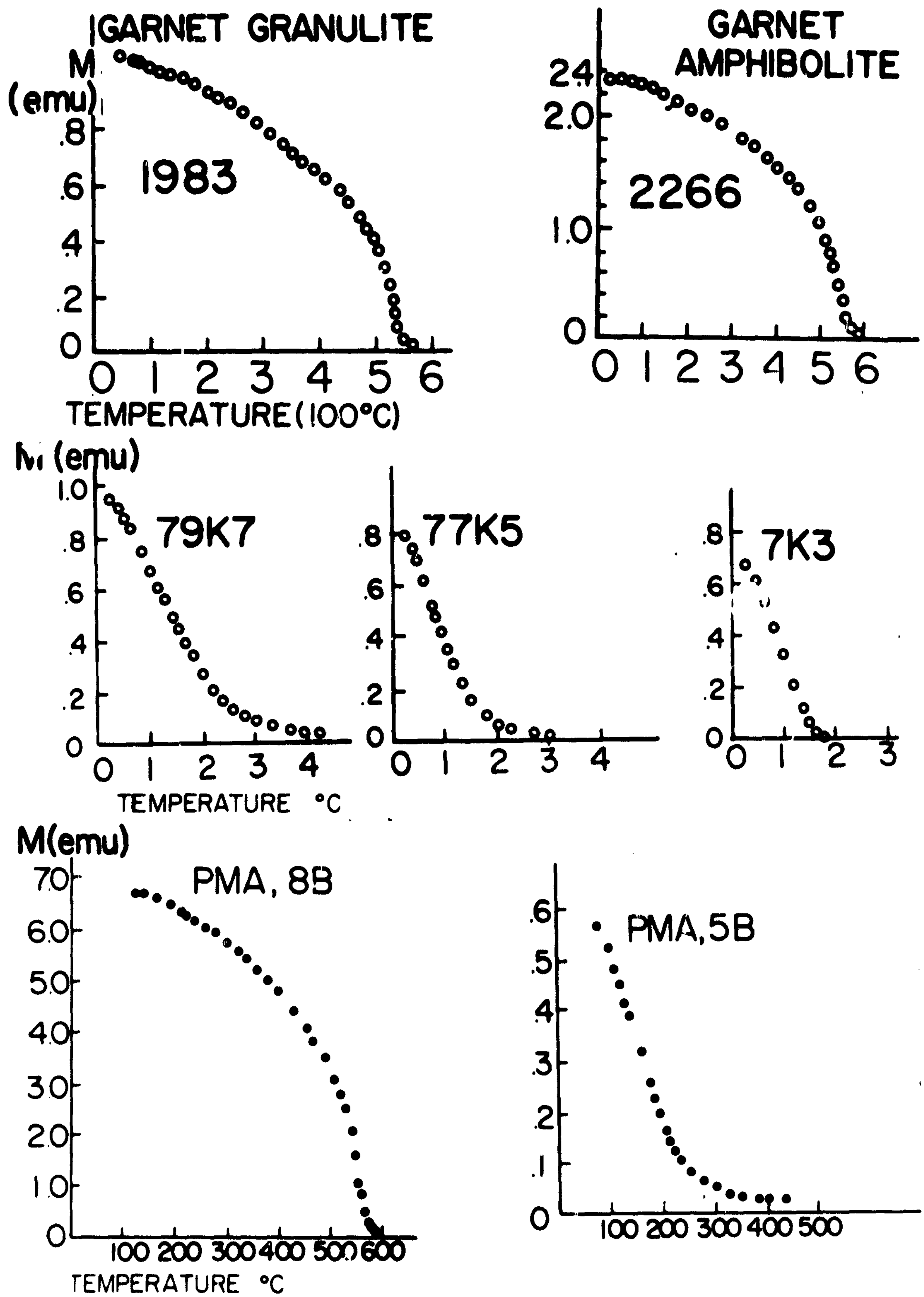
Shown in Figure 1 are Curie point curves for garnet granulite (1983) and garnet amphibolite (2266) from the Colorado plateau; pyroene granulties (79K7, 77K5, and 7K3) from Kilbourne Hole; and, an amphibolite grade rock from mid-crustal depth (PMA, 8B) and a lower crust granulite (PMA, 5B) from Potrillo maar.

All Colorado Plateau xenoliths have 550°C Curie points no matter what lithologies were evaluated. Kilbourne granulites have Curie points <300°C and Potrillo maar rocks have either 550°C Curie points (mid-crustal levels) or < 300°C Curie points (lower crust). The magnetic bottom beneath the Southern part of the rift should be no deeper than about 15 km as indicated by point A in Figure 2, which is the depth of the 550°C Curie point on a reasonable geotherm for the rift. If ilmenite dominates as is the case for Kilbourne Hole xenoliths then the magnetic bottom may be as shallow as 8-10 km (point B on Figure 2). Therefore, at best we may have only half the crustal thickness beneath the southern part of the rift made up of magnetic rocks.

Away from the central part of the rift where the geothermal gradient becomes more shallow, and the magnetic mineralogy may be developed in a more oxidizing environment, the effective magnetic crustal thickness should increase.

CRUSTAL MAGNETIZATION BENEATH THE RIO GRANDE RIFT

Wasilewski, P. J. and Padovani, E. R.



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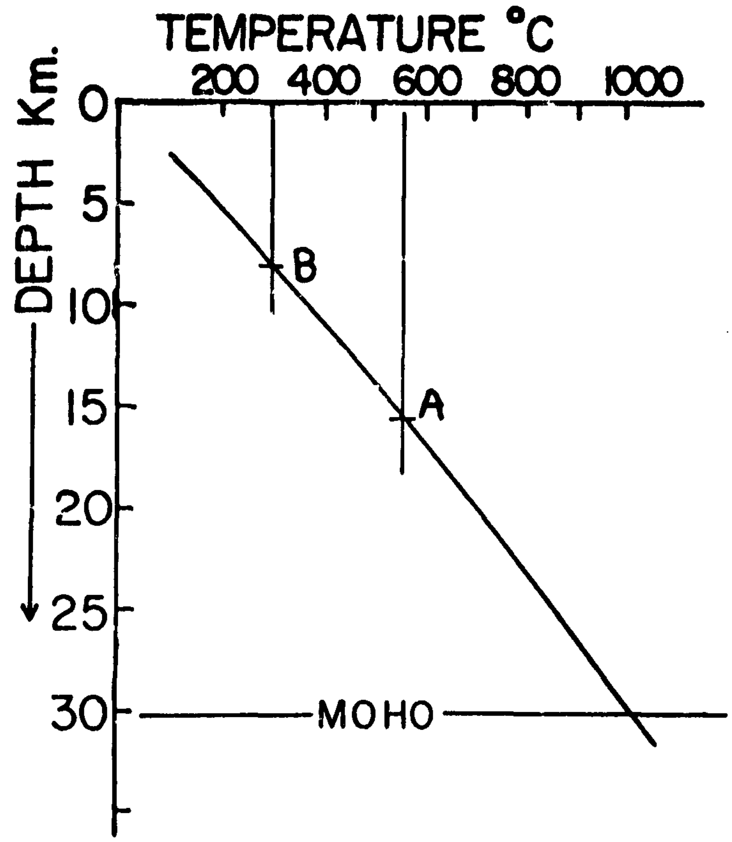

THE ROLE OF SEISMIC REFRACTION DATA FOR STUDIES OF THE ORIGIN AND EVOLUTION OF CONTINENTAL RIFTS

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Sengör and Burke (1978) and Baker and Morgan (1981) propose that there are two fundamental classes of continental rifts: active and passive. These two categories are hypothesized to arise from two different basic tectonic mechanisms, although a given rift system may combine a mixture of both. Active rifts are generated by deep-seated forces originating in the asthenosphere, i.e. mantle convection processes. Active rift systems are often extensions of the mid-ocean ridge/rift system, are usually strongly volcanic, and may include smaller scale "hot spots" and dome features. In passive rifts, the mobile asthenosphere plays a secondary role, "filling in" where the lithospheric plates are split or stretched thin by plate interaction stresses transmitted laterally from a distance. These two "end member" rift categories are thought to manifest the rifting-volcanism-uplift cycle in different sequences and in differing degrees of importance--leading to a wide variety of rift features unevenly distributed in space and in time of origin. (Sengör & Burke, 1978).

Whatever the basic driving forces, universal attributes of rifts are crustal extension and thinning and asthenospheric upwelling or modification of the uppermost mantle beneath the rift axis. Analysis of geological, geophysical and geochemical data from various rifts often suggests strong asymmetry in deep lithospheric structure and thermal properties with respect to the trend of the surface grabens. Recent researches have also hinted at considerable structural and thermal complexity within the crust, such as partial melt layers (Sanford et al., 1977), upper crustal low-velocity or ductile layers (Mueller, 1977), and anomalous midcrustal low electrical resistivity (Hermance and Pederson, 1980). The importance of these shallower features for the driving mechanisms and evolution of continental rifts is intriguing but speculative at present (Fuchs, 1974). Better knowledge of the seismic parameters--"layer" boundaries, compressional and shear velocity structure and anelastic attenuation (Q)--in the crust and uppermost mantle in rift zones are powerful constraints in deducing limits on deep composition, structure, stresses and thermal conditions, which are needed for realistic modeling of rift dynamics. Seismic refraction is the only geophysical technique capable of yielding high-precision velocity data below the Moho in order to adequately probe major continental rift systems at the required scale (lateral dimensions: 200 to 1000+ km). Other geophysical techniques--especially very high resolution but expensive (in terms of cost per linear unit of profile), "COCORP-like" vertical seismic profiling--are complementary to refraction profiling and need to be integrated with it for effective studies of the continental crust.

It has recently been recognized that the upper mantle is seismically anisotropic (Bamford and Crampin, 1977). While there is general agreement that mantle anisotropy probably results from the preferred orientation of olivine crystals, a variety of processes leading to lateral and vertical distributions of anisotropy have been proposed (Fuchs, 1977; Bamford, 1977). Measurement of anisotropy beneath continents promises to reveal much about the dynamics and petrology of the lithosphere-asthenosphere system, which is a region of prime interest for continental rifting processes. Indeed, anisotropy may help clarify some of the apparently conflicting Pn velocity

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evidence for or against the interpretation of a "cushion" or "pillow" of anomalous mantle material beneath the Rhinegraben rift system (Prodehl et al., 1976). Details of the lateral and vertical distribution of Pn anisotropy can in principle provide clues as to deformations and/or stress history at the asthenosphere/lithosphere boundary (Bamford et al., 1979). For example, deformation ocurring in cold lithosphere would induce a depth-independent anisotropy characteristic of the long-term stress pattern. On the other hand, mineral alignments could be induced at the lithosphere/asthenosphere boundary by "global" or by more localized tectonic events and then frozen in as the lithosphere cools--this would lead to depth dependent anisotropy indicative of stress history. In practice, the separation of the effects of velocity anisotropy from those of lateral variations in both structure and velocity requires a large seismic network of intersecting profiles and fans. For the continents, only the Rhinegraben area of central Europe presently approaches a sufficient coverage by many refraction profiles to permit analyses for anisotropy. (Anisotropy studies under the much thinner oceanic lithosphere are logistically much easier to carry out.)

In recent years, significant increases in speed and capacity of digital computers has lead to analysis techniques that permit much more sophisticated and detailed earth models to be derived from amplitude and wave form data as well as from travel times. The most important of these are:

- Synthetic seismogram modeling extracts a maximum of information from amplitude and waveform data. Particularly useful are techniques such as the modified reflectivity method (Kind, 1978), which are capable of calculating "complete" seismograms (i.e., both body and surface waves) with few restrictive assumptions. Examples of synthetic record sections will be presented that illustrate effects of assumed structural variations (velocity gradients, laminated interfaces, etc.) on seismogram characteristrics.
- Optimized travel-time inversion and ray tracing programs allow modeling of lateral heterogenieties.
- Time-Term analysis (Bamford, 1976) helps define refractor contours and velocity anisotropy for large data sets.
- Joint inversion techniques incorporating other geophysical data sets (gravity, Curie isotherms, electrical resistivity etc.) together with seismic velocity and Q models help select more physically consistent models.

For studies of continental lithosphere in general, and continental rifts in particular, there are three main requirements for seismic refraction profiles that must be met to achieve significant advances in resolution and detail over past data:

(1) Relatively close station spacing (1 to 5 km) over ranges of 30 to 300 km. Even greater distances of (1000 to 2000 km) are required if information is needed about deep lithospheric and asthenospheric structure. This implies 100 or more instrument stations along major profiles.

(2) Energy sources (usually explosions) of sufficient power are needed to penetrate depths of at least 50 km. Unless convenient lakes or other water bodies are available to optimize coupling efficiency, large explosions (many tons) may be needed for continental profiles. This can be a severe cost and logistic restriction, which is compounded by the fact that contemporary continental rift systems occur in tectonic terrains where seismic attenuation (1/Q) of crustal materials is large.

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(3) A carefully selected set of reversed profiles at several azimuths (both axial and crosslines for rifts) are required to resolve large-scale lateral variations and to reduce uncertainty in determination of "layer" velocities. A systematic long-term program of many intersecting profiles is required to obtain sufficient data for anisotropy studies.

Present status of refraction data and interpretation for four major Cenozoic continental rifts:

Rhinegraben. The classical example of a rift valley, it is undoubtedly the best studied with more than 25 published profiles--many of them reversed (Prodehl et al., 1976). Detailed work began in 1966 using available quarry blasts for sources. These sources were often at unfavorable locations or of insufficient energy to permit layout of profiles best suited to clarify tectonic problems. This early work (mostly crosslines) lead to the concept of a low velocity (Pn velocity ~7.6 to 7.7 km/s), "lens" or "cushion" of "anomalous mantle" material separating the bottom of the thinned crust (~25 km depth) from normal mantle (Vp ~8.0 to 8.2 km/s). The "anomalous mantle" was believed to be the seat of the fundamental driving force for rifting in the Rhinegraben. Subsequent reversed, axial profiling forced abandonment of the rift cushion concept to the extent that the true Pn velocities were found to be ~8.0 km/s but the crust-mantle boundary depth was not greatly changed. The early apparently anomalous Pn velocity ~7.6 km/s is now explained in terms of a ~8.0 km/s refractor of varying topography (Prodehl et al., 1976). Another important result of seismic refraction interpretation is that outside the Rhinegraben, the crust-mantle transition (Moho) is a sharp first-order discontinuity whereas inside, the transition is a (? laminated) gradient zone about 5 km thick. Many record sections in the Rhinegraben suggest evidence for a silicic low velocity or ductile layer near the base of the upper crust (Prodehl et al., 1976; Mueller, 1977), which Fuchs (1974) has discussed as of possible importance as a driving mechanism for graben formation. Seismic profiles are dense enough so that anisotropy can be observed but the significance is still debated. Modern analysis techniques (synthetic seismograms, ray tracing, etc.) have been extensively applied to Rhinegraben seismic data.

Rio Grande Rift (RGR). As of 1981 only three refraction profiles longer than 100 km penetrated the Moho: One N-S axial profile (Olsen et al., 1979), a parallel one largely outside the western rift margin and a E-W crossline (Cook et al., 1979) at the southern end. All of these profiles are unreversed but there are several deep vertical (COCORP) reflection profiles (Brown et al., 1979), which were jointly used in interpretation. Several axial and crossline profiles are to be reversed and extended in September, 1981. Existing profiles indicate moderate crustal thinning (~33 km beneath rift, $\geq$40 km under flanks) and a Pn velocity ~7.6 km/s uncorrected for possible dip. Unreversed crossline data suggest a "rift cushion" of ~7.4 km/s may exist beneath the southern rift (Cook et al., 1979). Keeping in mind the Rhinegraben experience, the cushion interpretation may need revision in favor of lateral structural variations and/or anisotropy effects. Of great interest in the RGR is the presence and possible extent of a (? thin) sill-like magma body or partial melt layer sharply separating the upper and lower crust at a depth of 18 to 20 km (Sanford et al., 1977). This layer is well observed by several geophysical methods in the central rift (Socorro area). Strong undercritical seismic reflections at wide incidence angles and other indicators suggest the layer may extend (? intermittently) beneath other parts of the RGR. Current profiling is aimed at extending knowledge of this (? unusual) mid-crustal feature. Correlation of this low-rigidity layer with the "silicic ductile layer" (Fuchs, 1974) at similar depths in the Rhinegraben is presently very uncertain. Modern synthetic seismogram and ray tracing modeling is being applied to RGR data.

East African Rifts. A sparse (average station separation ~30 km) 360 km refraction line along the approximate axis of the eastern branch (Gregory rift) of the East African rift system has been discussed by Grifiths et al., (1971). Interpretation in terms of travel-times is very uncertain but suggests a transition (? Moho) from ~6.4 km/s crust to ~7.4 km/s (? mantle) material at ~20-km depth. Five 120- to 250-km long reversed profiles (station interval 5 to 10 km) have been obtained in the Afar region (Berkhemer et al., 1975). Anomalous mantle with Vp = 7.3 to 7.6 km/s, upper surface varying in depth between 16 km (north) and 26 km (south) and of thickness of 15 to 40 km is suggested. Berkhemer et al. applied travel time correlation, time-term analysis and amplitude analysis in their interpretation.

Baikal Rift. Probably the most seismically active of the four major rifts. Published interpretations (Puzyrev et al., 1978) indicate the main rift area underlays Lake Baikal and is the western boundary of an extensive zone of crustal extension similar in several respects to the Basin and Range province of the western United States. Both axial and crossline profiles (including some very long DSS profiles) overlap Lake Baikal and the trans-Baikal extension zone. Anomalous mantle (Vp = 7.7 km/s) extending from depths between 30 and 60 km and a low velocity zone (Vp ~ 6.2 km/s) in the uper crust are dominant features of the published interpretations. Unfortunately, almost no record sections of the Baikal region have been published in standard Soviet journals, so additional reinterpretation using synthetic seismogram modeling has not been performed to my knowledge.

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DEEP STRUCTURE OF THE MISSISSIPPI EMBAYMENT FROM SEISMIC REFRACTION MEASUREMENTS; D. Peters, W. Lutter, A. Ginzburg*, A. Walter, and W. Mooney, U.S. Geological Survey, Menlo Park, CA 94025, (*also at University of Tel Aviv, Israel)

Considerable attention has been focused on the Mississippi embayment (figure 1) due to its seismic activity which is highlighted by the destructive New Madrid earthquakes of 1811 and 1812. Attempts to understand the cause of this activity have led to considerable speculation regarding the crustal structure of the embayment. Previous work by Erwin and McGinnis (1975), Hamilton and Russ (1980), and Zoback and others (1980) suggests that present-day seismicity may be related to the reactivation of an ancient rift zone. This study presents new seismic refraction data concerning the crustal structure of the embayment.

In September 1980 the U.S. Geological Survey recorded 34 shots from nine shot points in the northern Mississippi embayment (figure 2). One hundred seismographs recorded each shot in profiles parallel and perpendicular to a northeast-southwest-trending aeromagnetic anomaly along the axis of the embayment. The interpretation of the profiles shows the presence of the following velocity sequence: 1.95, 6.05, 5.6, 6.28, 6.75, 7.3, and 8.0 km/s.

Geologically, the velocity layering agrees well with known embayment structures. The 1.95 km/s layer represents Cenozoic and Mesozoic unconsolidated sediments and is up to 0.75 km thick. These sediments represent embayment subsidence and sedimentation in a syndepositional basin setting, and were deposited over an erosional surface of Paleozoic rocks. The unconformity which exists at this Mesozoic-Paleozoic boundary is clearly evident by the sharp velocity contrast of the 1.95 and 6.05 km/s layers. The 6.05 km/s layer probably represents an Early Paleozoic carbonate sequence which overlies a Late Precambrian-Early Paleozoic sandstone sequence represented by the 5.6 km/s layer. This layer constitutes a 5 km-thick low velocity zone which is apparent on all seismic sections as a cut-off of the 6.05 km/s arrivals causing a shadow zone 20-30 km in length.

The 6.28 km/s velocity layer is interpreted as undifferentiated igneous basement, including older Precambrian rocks and younger late Precambrian intrusive and extrusive igneous rocks underlying the Paleozoic sediments. The 6.6 km/s layer represents mafic or basaltic lower crust. The 7.3 km/s layer is that portion of the lowermost crust that has been intruded by mantle material, thus leaving only $\sim$5 km of unaltered (6.6 km/s) lower crust in the direction of the axial profile. This intrusion represents the ancient rifting and mantle upwelling in the Precambrian as suggested by Erwin and McGinnis (1975).

The 7.3 km/s layer may not be present in the southern portion of the embayment. The 8.0 km/s velocity zone represents mantle material.

The Mississippi embayment shows similarities in crustal structure to other continental rift systems (figure 3). An anomalous, high velocity zone represented by the 7.3 km/s layer in the embayment has been observed in the Jordan-Dead Sea rift system (Ginzburg and others, 1979) and the Limagne graben (France) (Hirn and Perrier, 1974). Updoming of the crust-mantle boundary and injection of upper mantle material into the lower crust has been suggested by Fuchs and others (1980) for the Rhinegraben rift. The structure of the embayment must be studied in greater detail to determine whether realistic comparisons can be drawn, and to what degree of reliability general crustal models can be applied.

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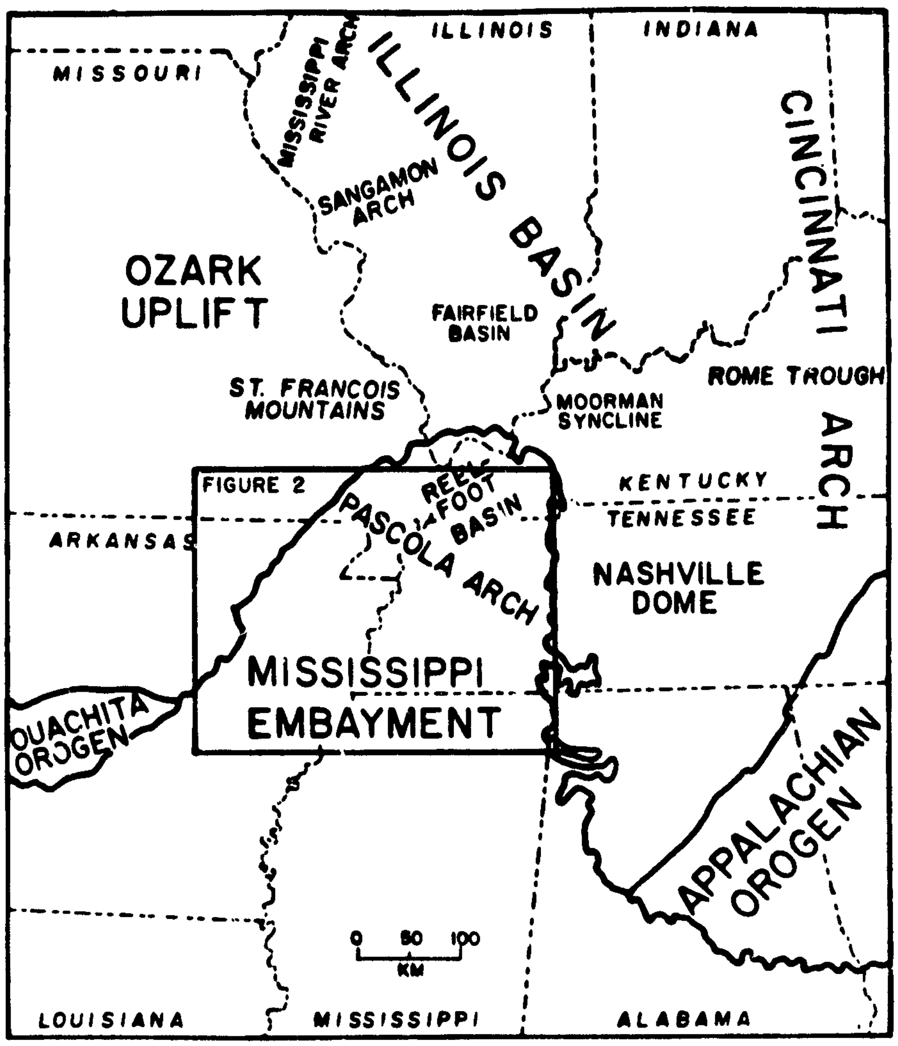

Figure 1

General location map of the Mississippi Embayment of the Central U.S.

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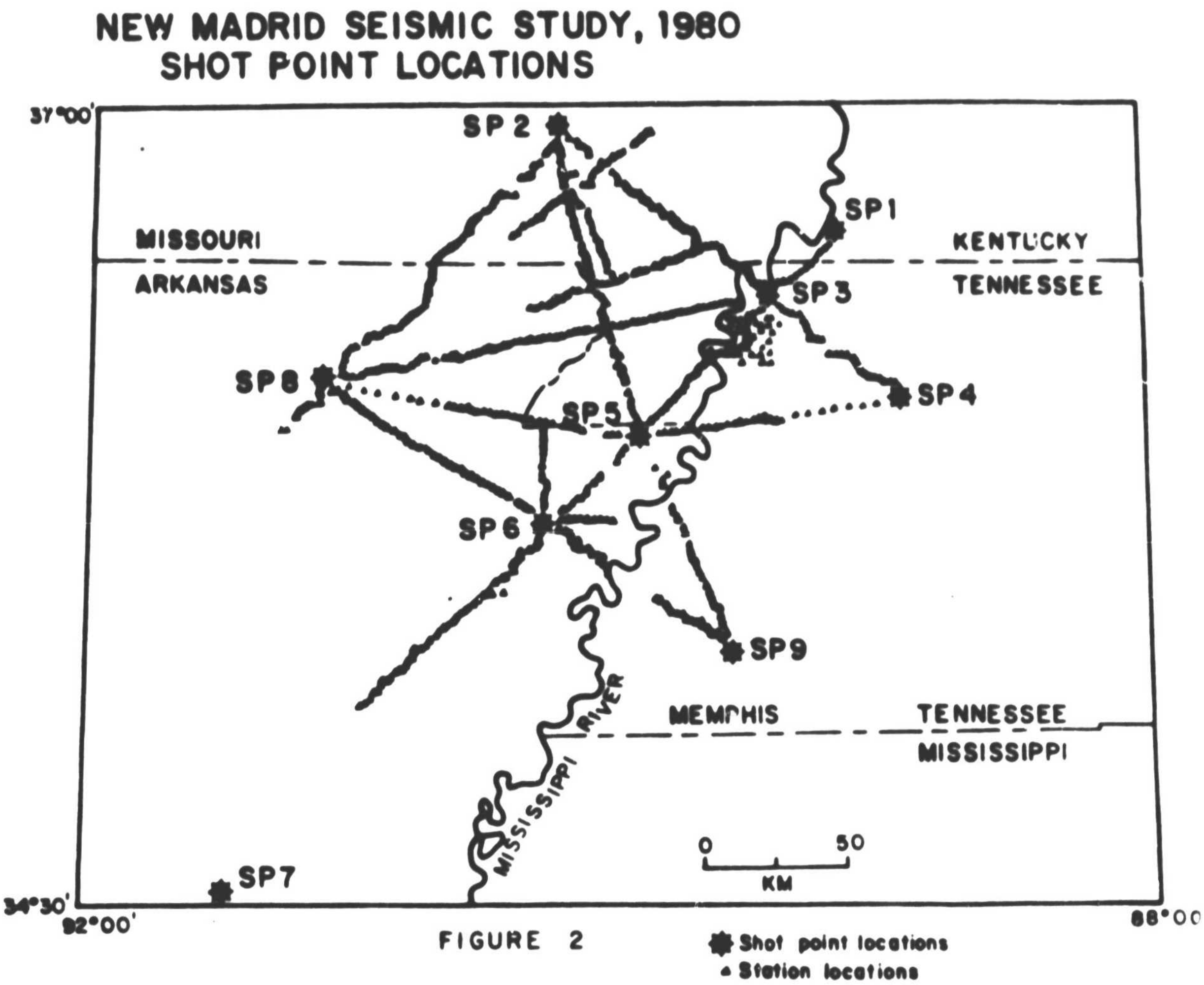

FIGURE 2

Shot point and recording station location map for the seismic refraction profiles.

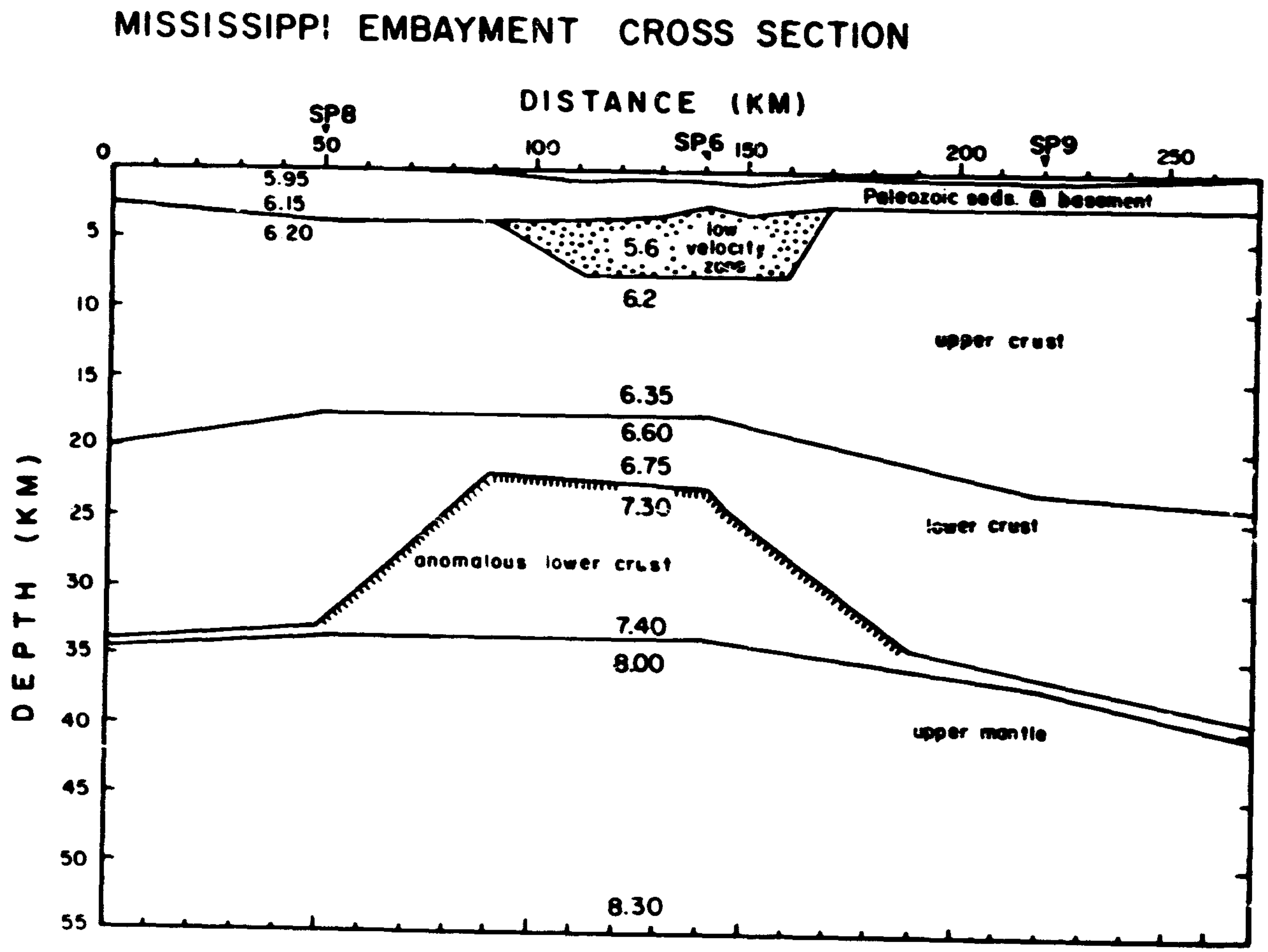

FIGURE 3

Preliminary crustal cross-section across Mississippi Embayment based on an intertretation of the 1980 USGS seismic refraction profile S.P. 8 - 6 - 9 (Figure 2).

CONSTRAINTS ON RIFT THERMAL PROCESSES FROM HEAT FLOW AND UPLIFT DATA.
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Heat flow is fundamental to the understanding of the genesis of continental rift zones. Active thermal processes are dramatically manifested by volcanism in rift zones, and temperature is perhaps the most important parameter controlling the processes of rifting. Surface heat flow measurements provide information from which deductions can be made on the nature, magnitude, and history of the thermal processes and subsurface temperatures in rift zones. Uplift in rift zones is an isostatic response to density changes and thinning of the lithosphere, which result from structural, chemical and thermal modification of the lithosphere during rifting. If the structurally and chemically induced components of uplift can be identified, therefore, the thermal component of uplift can be used to study thermal modification of the lithosphere during rifting.

Published heat flow data are available from five Cenozoic continental rift systems, and the means of the data from the floors of the rifts are generally remarkably similar: Baikal rift, 97±22 mW/m**2 (n=52) (Lubimova *et al.*, 1972); U.S. Basin and Range province, 70-125 mW/m**2 (Lachenbruch and Sass, 1978; Blackwell, 1978); Kenya rift, 98±48 mW/m**2 (n=7) (Morgan and Wheildon, 1981); Rhinegraben, 112±34 mW/m**2 (n=8) (Bram, 1979); and Rio Grande rift, 107±27 (n=25) (Reiter *et al.*, 1979). Contrasting data have been published for the lakes of the western rift of the East African rift system: Lake Malawi, northern, 22±13 mW/m**2 (n=12), central, 96±25 mW/m**2 (n=5), southern, 31±3 mW/m**2 (n=3) (Von Herzen and Vacquier, 1967); Lake Tanganyika, all data except one high value of 150 mW/m**2, 38±13 mW/m**2 (n=11) (Degens *et al.*, 1971), Lake Kivu, 17-185 mW/m**2 (Degens *et al.*, 1973). Data from the Basin and Range province are quoted as a range because heat flow subprovinces have been identified within the physiographic province. Lake Kivu data are also given as a range representing the wide scatter in the results from five measurement sites. In general the heat flow data indicate high heat flow, 70-125 mW/m**2, from the floors of the rift valleys, with the exception of parts of the western rift of the East African rift system.

The significance of the similarity between the high mean heat flow values from the various rifts is not well defined by the available data sets. Heat flow data usually represent only the conducted portion of the heat loss, and in regions of active tectonics, convection by groundwater and/or magma can be significant, and should be included if a complete thermal analysis is to be attempted (e.g. see Blackwell, 1978). Heat flow values in the range 70-125 W/m**2 over large areas indicate temperatures near the melting point in the lower crust and upper mantle, and perhaps this has a self-regulating effect on high heat flow on a province wide scale in the continental crust. Higher heat flow causes upward migration of magma to shallower depths where cooling is more rapid. Very high heat flow values have frequently been reported associated with young magmatic activity, but are generally, perhaps incorrectly ignored in regional heat flow analyses. Lower heat flow reduces the convective heat losses. and the anomaly cools slowly by conduction alone. The Basin and Range province is the only rift system for which a reduced (lower crustal/upper mantle) heat flow value has been determined from a heat flow vs. heat production plot. Lachenbruch and Sass (1978) indicate that this determination is meaningless if data from all areas are taken, but Blackwell (1978) shows that the determination can be made if data are used only from areas where provincial volcanism is older than 17 m.y. The Basin and Range reduced heat flow value of 60 mW/m**2 would explain surface heat flow values of 65-100 mW/m**2 in the other rifts with a steady-state temperature profile intercepting the melting point curve in the upper mantle. Alternatively the rift heat flow values in this range could include a transient component of heating caused by magmatic intrusion into the crust. Higher heat flow values indicate some transient component of crustal heating.

If the data from the lakes in the western rift of the East African rift system are accepted to be reliable, the contrasting flow heat low values from Lakes Malawi and Tanganyika provide a clue to the source of the thermal anomalies in the East African rift system. The high heat flow values from both these lakes are located near structural shifts in the lakes, and may be caused by fault-controlled groundwater convection (Von Herzen and Vacquier, 1967; Degens et al., 1971). Regional low heat flow is indicated. This low heat flow corresponds to sections of the rift system with little or no Tertiary-Recent volcanism. The sections of high heat flow, the Kenya rift and Lake Kivu are areas of extensive Tertiary-Recent volcanism (e.g. see King, 1970). The immediate source of the high heat flow in the East African rift system would therefore appear to be crustal magmatic convection, related to the volcanic activity.

It is possible to model the heat flow data across the Kenya rift using a dike intrusion model into the axial portion of the rifted crust (Morgan, 1973), and a similar model has been proposed by Zorin (1981) to explain heat flow data across the Baikal rift. In both the Kenya and the Baikal rifts, the absence of broad regional heat flow anomalies outside the floors of the rifts indicates that conducted thermal anomalies from the mantle have not yet reached the surface. For the Basin and Range province and western U.S. the thermal history is much more complex, and the high heat flow has been interpreted as both the effect of convection during rifting and extension (Lachenbruch and Sass, 1978), and as the cooling of a widespread thermal event (Blackwell, 1978). One of the major problems in separating these different thermal models is that the thermal histories of the rifts are unknown.

One method of attempting to unravel the thermal history of an area is to use relationships between heat flow and elevation in a similar manner to which heat flow and bathymetry have been linked to the age of the ocean floor (e.g. see Parsons and Sclater, 1977). This technique has been successfully applied to the rift-like Snake River Plain to predict the heat flow and surface elevation of the plain after the passage of a hot spot currently under Yellowstone (Brott et al., 1978, 1981). With many of the rifts, however, the problem is in reverse, subsidence has not yet started, and a thermal model is needed to match uplift rates to the surface heat flow data. Thermal processes in the lithosphere can cause uplift by two mechanisms: i) thermal expansion of the lithosphere, and ii) thermal thinning of the lithosphere (conversion of lithosphere to asthenosphere). The first mechanism has most recently been modelled by Mareschal (1981), but does not readily explain the relatively high rates of uplift (0.1 km/m.y. and greater) that are indicated by a variety of data sets from some rifts and continental uplifts. A simple model for the thermal thinning of the lithosphere is therefore presented.

A thermal source is assumed at the base of the lithosphere. The origin of this source is irrelevant to the model, but it is probably derived from convection in the asthenosphere driven by thermal and/or chemical instabilities. To keep the model simple a continuous point source model is assumed (Carslaw and Jaeger, 1959, p.261) with a strength such that it would increase the surface heat flow from 40 to 80 mW/m**2 if placed at the base of a 40km thick crust. The lithosphere is assumed to start with a shield geotherm and the base of the lithosphere is assumed to be the pyrolite solidus. The heat source is assumed to ascend with the lithosphere/asthenosphere boundary, and for this simple model this ascent was effected in 5 km increments, the time for each increment being determined by time required for the temperature to rise to the pyrolite solidus 5 km above the source. The smoothed ascent of the base of the lithosphere from 200 to 40 km predicted by this model is shown in Figure 1, and has a mean rate of 4.3 km/m.y., starting at over 12 km/m.y. at 200 km, and slowing to 3 km/m.y. at 40 km. Different assumptions about the source strength, starting geotherm, solidus curve and ascent increment change these rates, but the significant result is that with reasonable input parameters, the lithosphere thins at a rate much faster than a conducted thermal perturbation would propagate through the lithosphere. The lithosphere is essentially unheated in advance of its rising base. More realistic source configurations and variable source strengths would make the thinning more complex, and perhaps episodic, but the mean rate would be of the same order as predicted by the simple model.

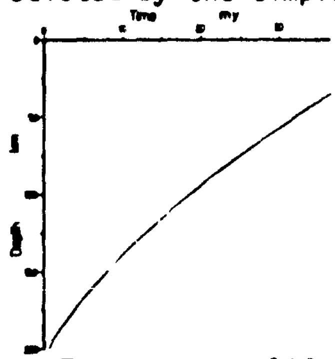

Figure 1. Smoothed ascent of the base of the lithosphere based on an initial shield geotherm with a point source moved upwards in 5 km increments as the temperature of the pyrolite solidus is reached.

To compute uplift from lithospheric thinning, the model of the lithosphere floating in the asthenosphere given by Crough and Thompson (1976) has been used. From this model it can be shown that lithospheric thinning, dL, is related to surface uplift, U, by $dL = U*Ra/(Rl-Ra)$, where Ra and Rl are the densities of the asthenosphere and mantle lithosphere respectively. Assuming $Ra = 3.2$ g/cm**3 and $(Rl-Ra) = 0.05$ g/cm**3, a lithospheric thinning of 64 km results in 1 km of uplift, which translates to an uplift rate of 0.1 km/m.y. for a thinning rate of 6.4 km/m.y.

The thermal thinning model for the lithosphere therefore predicts a rate of uplift of the same order as the rift data suggest. The other implications of the model are that conducted broad thermal anomalies are not to be expected over young rift systems, although more local anomalies may be produced by penetrative convection of magmas related to the thinning into the crust. The model also predicts a progressive shallowing of the thermal anomaly which would result in a progressive shallowing of the minimum depth of magma generation associated with rifting through time. A progression of this type has been deduced in East Africa by Wendlandt (this volume). Broad regional heat flow anomalies are only predicted for rift systems where the lithosphere has been considerably thinned for a period of time greater than the time constant of the remaining lithosphere, several millions of years.

The model presented is strictly applicable only to active rifting, where the rifting is caused by an upwelling of asthenospheric material. In passive rifting, produced in response to lateral stresses resulting from plate interactions, lithospheric thinning can be caused by extension. This extension can result in an upwarp of the isotherms, but as more active convective processes are often also evident from volcanism, it is possible that secondary convection in the asthenosphere may result in further lithospheric thinning with results similar to those produced by the model. To rigorously test these hypotheses, however, reliable data on heat loss and uplift rates are required from all active rift zones.

Acknowledgements. This work was funded in part by grant nos. EAR77-23354 and INT78-16649 from the Earth Sciences Program and Office of International Programs of the National Science Foundation respectively. A portion of this research was performed at the Lunar and Planetary Institute which is operated by the Universities Space Research Association under contract no. NASW-3389 from the National Aeronautics and Space Administration.

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THE RELEVANCE OF INTRAPLATE VOLCANIC CENTRES TO RIFTINC: AN EXAMPLE FROM JEBEL MARRA, WESTERN SUDAN; **P.M. Bermingham, J.D. Fairhead and G.W. Stuart, Dept. of Earth Sciences, University of Leeds, Leeds LS2 9JT, UK.**

The intraplate volcano of Jebel Marra in western Sudan is the furthest east of a line of Cainozoic volcanic centre in N. Africa which are associated with domal uplift and are linked by a zone of negative Bouguer gravity (Fairhead, 1979). These domal uplifts and their associated volcanism may represent an earlier stage, in the processes that could eventually lead to continental disruption, than that represented by the East African Rift System. Thus to determine the lithospheric structure associated with these centres a detailed geophysical and geochemical study of the Jebel Marra volcanic province has been initiated by the University of Leeds. This contribution is restricted to a preliminary interpretation of the geophysical data.

A total of 860 gravity measurements were made in western Sudan between March 1980 and May 1981 using a LaCoste Romberg gravity meter. Measurements were taken at about 10 km intervals along tracks around Jebel Marra decreasing to 3 km intervals over the volcanic centre itself. This network of gravity measurements was tied to 3 existing gravity base stations at Zalingei, Nyala and El Fasher (Fig. 1) which were in turn tied to the IGSN71 via the Khartoum gravity base (Isaev and Mitwalli, 1974). Heights of stations were obtained using standard barometric techniques tying measurements of Trigometric points and Bench marks. Locations were determined using existing maps, ERTS imagery, a satellite navigation system (SatNav 801) and astronomical observations. The preliminary Bouguer gravity and geology map is shown in Fig. 1. The map clearly indicates that the surface volcanics are stricted to a very small central part of the domal uplift which is ~500 km in diameter and is delineated by the -60 mGal contour. The amplitude of the negative Bouguer anomaly associated with the basement uplift is about 60 mGal (ie -50 to -110 mGal). Over the volcanic area a smaller wavelength negative Bouguer anomaly exists with amplitude 30 mGal. (i.e. -110 to -140 mGal). This anomaly is caused by the thick pile of low density pyroclastics (see profile). Negative gravity anomalies as well as crustal uplift are also associated with the Meidob volcanic centre 250 km NE of Jebel Marra indicating the domal uplift is elongated in a NE direction. Detailed mapping of this area was not possible due to logistic difficulties.

The cause of the long wavelength Bouguer anomaly is clearly subcrustal and is most probably due to a hot low density low seismic velocity and low Q region within the upper mantle similar to that modelled beneath the East African Rift System. To reduce the non-uniqueness of the gravity models for this body, seismological data have been collected. A mobile, 4 element SPz, seismic array with station separation of 10 km was located on basement at 2 sites close to the centre of the uplift and 1 site off the main anomaly close to Nyala town. These 3 sites are shown in Fig. 1 and were operated for a total of 6 months. In addition a base reference seismic station was operated in Nyala throughout the experiment to enable relative P-wave delay times to be determined between array sites.

GEOPHYSICAL STUDY OF INTRAPLATE VOLCANO JEBEL MARRA

Bermingham, P.M. et al.

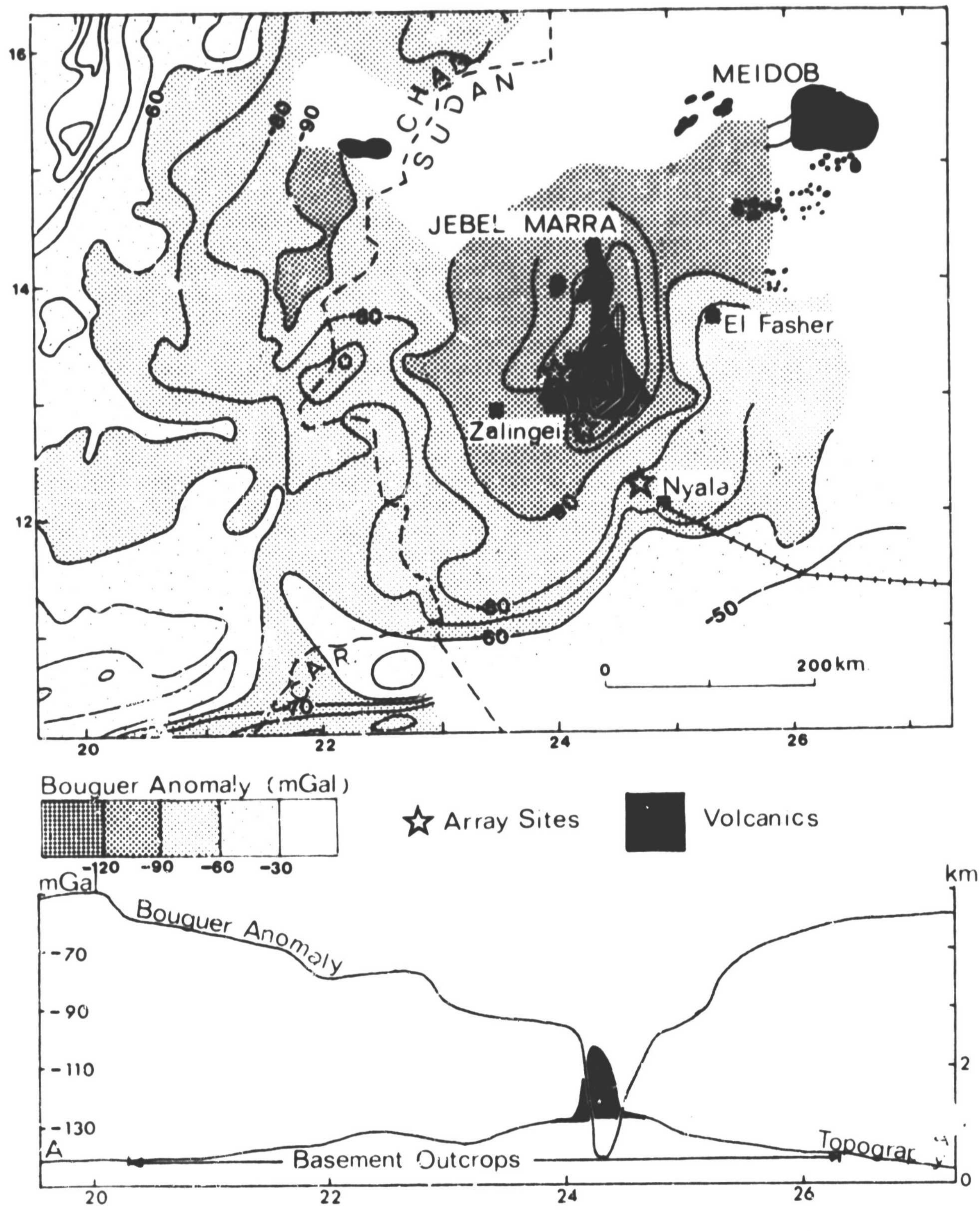

FIGURE 1.

GEOPHYSICAL STUDY OF INTRAPLATE VOLCANO JEBEL MARRA

Bermingham, P.M. et al.

Azimuthal variation of teleseismic P-wave delay and slowness anomalies and preliminary 3D interpretation of the gravity data will be presented and discussed. The regional setting of Jebel Marra within the framework of continental rifting will be discussed in the contribution by Fairhead and Browne(1981).

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THERMAL MAGNITUDES AND PERIODICITIES OF THE EVOLVING RIFT BOUNDARIES: STEADY STATE OR NOT? Kathleen Crane and Suzanne O'Connell, Lamont-Doherty Geological Observatory, Palisades, NY 10964

In the last ten years detailed observations along spreading centers have shown that the volcanic and tectonic processes are not steady state phenomena but vary with space and time along the strike of the accreting plate margins. Most dramatic were the discoveries of ephemeral hydrothermal fields localized over near surface magma chambers. The fields are limited in space as well as time yielding to the cooling processes of the lithosphere. Associated with these active vents are fresh volcanic sheet flows contrasting with the pillow ridges so ubiquitous on many of the cooler accreting margins. In addition, tectonic variability (zones of intense fissuring, faulting, up to the larger scale wide U-shaped vs. narrow V-shaped rift valleys) was discovered. The theories of along strike volcanic and tectonic variability and propagation evolved from the surveys at the Galapagos Rift, the Mid-Atlantic Ridge and E.Pacific Rise.

What controls the variability of axial width, depth, and structure? Are the presence of periodically spaced rise crest domes an indicator of recent subjacent volcanic activity which propagates down the strike of the rift? The thermal tectonic models which we should like to test are based on: 1) the evolution of volcanic cells between two cold boundaries (the bounding transform faults in the oceanic case), 2) propagation of these volcanic centers with time, 3) cyclic variability in rift valley width ranging from narrow to wide over a period of 1 m.y., and 4) the wavelength and variability of hydrothermal and volcanic activity depending on lithospheric thickness and the length of the subjacent heat source.

To test these models we use three active mid-ocean rifts and compare their morphologies and heat flow to the East African Rift - a well developed continental rift that grades from oceanic crust in the Afar to pseudo-continental crust with distance down the rift.

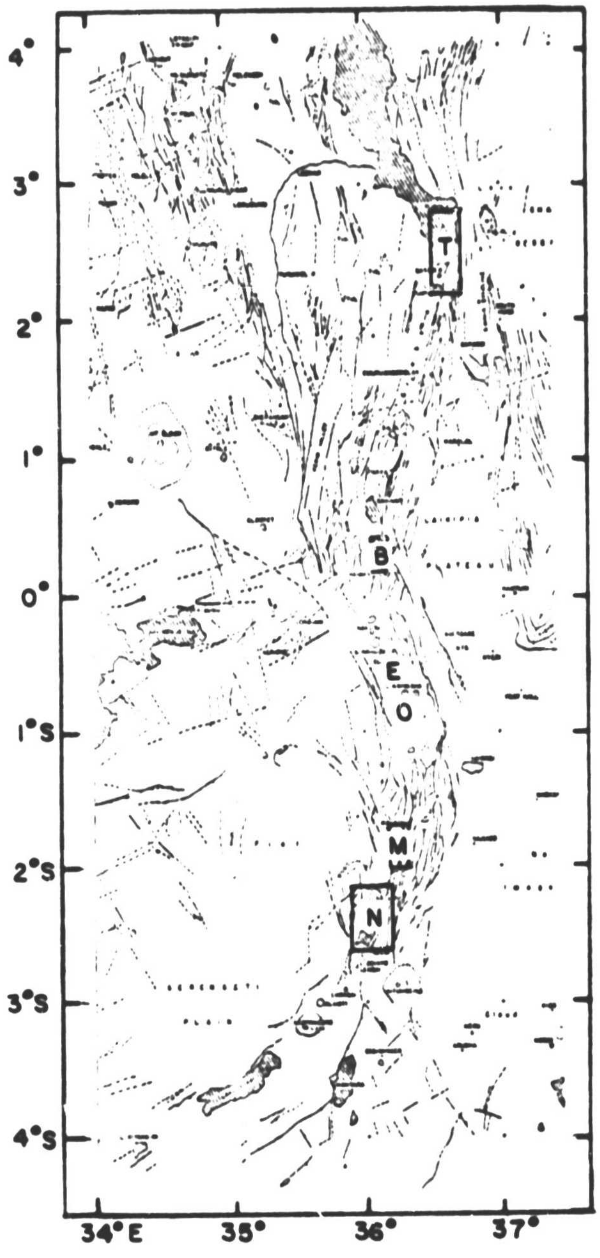
Extensive hydrothermal activity on the East Pacific Rise at 21^{o}N emits enough heat to cool approximately 36-72 km of the ridge crest out to a distance of 30 km from the accretionary boundary. Large morpho-tectonic domes on the EPR are spaced 40-80 km from each other. This may imply that major geothermal fields as well as major volcanic features are a periodic feature on the active spreading center. This is the case along the East African Rift where volcano spacing is nearly a linear function with crustal thickness. However, not all volcanoes are active within the rift. Average spacings of active major geothermal fields along the East African Rift in Kenya are $\approx$300km compared to the proposed 80 km for the East Pacific Rise. Convective heat loss along the Kenyan section of the East African Rift averages 17-46 MW per kilometer of rift length whereas at the East Pacific Rise 40-80 MW/km is calculated.

THERMAL MAGNITUDES AND PERIODICITIES

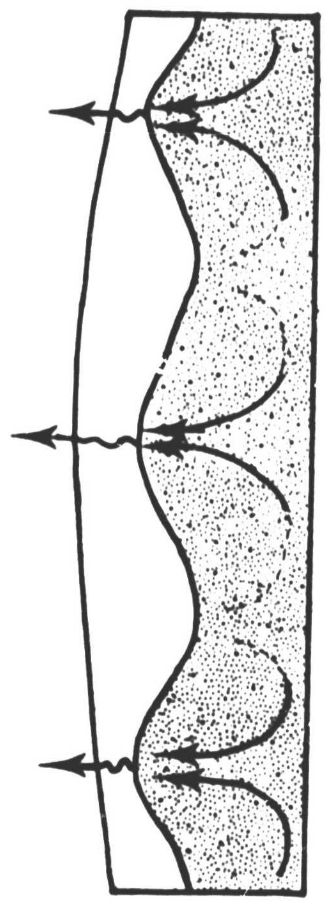
Crane, K. and O'Connell, S.

Morphotectonic information on older oceanic crust infers that major volcanic centers have propagated with time down the rift at a time integrated rate of many cm. per year.

From these relation we can postulate about the thermal evolution of accreting plate boundaries by mapping the frequency of major volcanic domes at discrete age intervals of the crust.



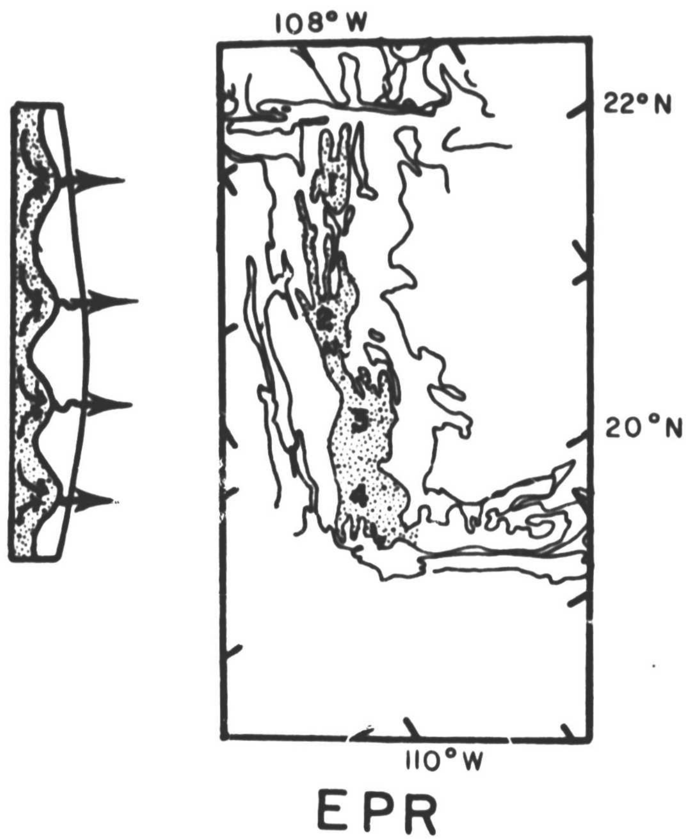
GREGORY RIFT

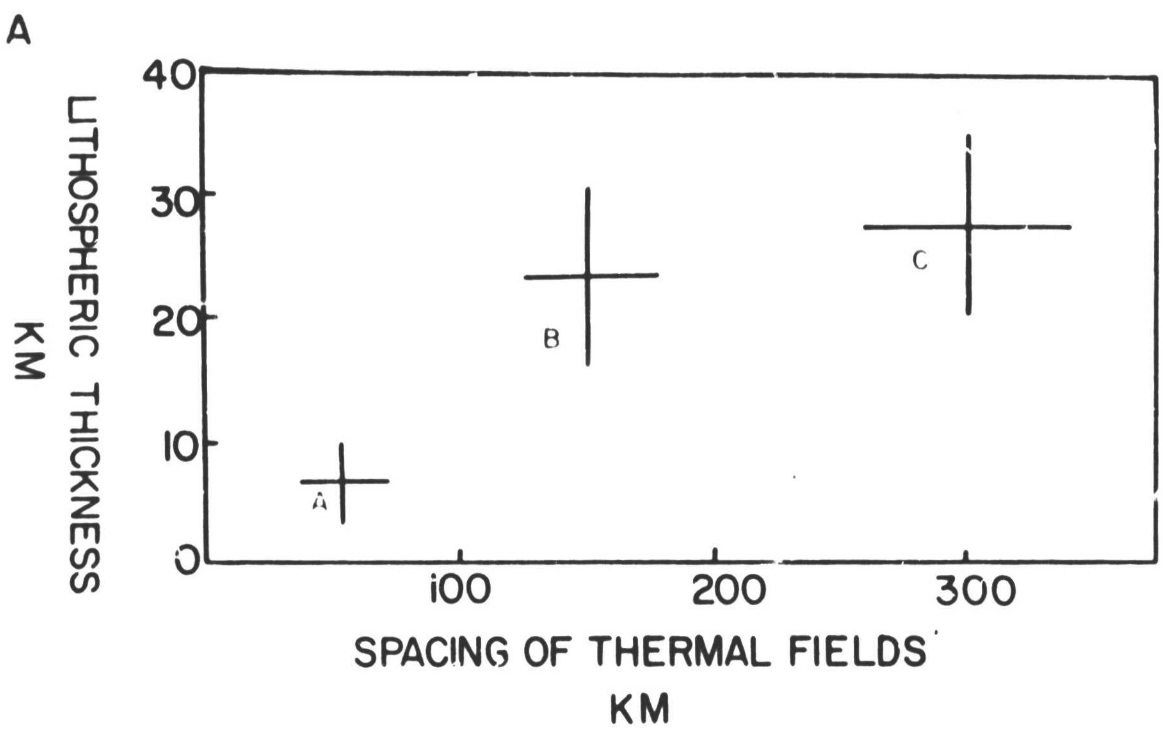


(Figure published courtesy of Elsevier Scientific Publishing Company)

THERMAL MAGNITUDES AND PERIODICITIES

Crane, K. and O'Connell, S.





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SATELLITE GEOPOTENTIAL ANOMALIES AND CONTINENTAL RIFTS. Herbert Frey and Michal Ruder, Geophysics Branch, NASA Goddard Space Flight Center, Greenbelt, MD 20771

Satellite geopotential data provide broadscale information on the structure and composition of the global crust, and may be useful in comparative studies of largescale tectonic structures. While continental rifts are generally narrow compared with the presently available resolution in satellite magnetic and gravity anomaly data, the disturbed region within which such rifts lie are of a scale where the satellite data can be important. These data are especially useful in that they provide global, three-dimensional constraints on crust and upper mantle structure, which is particularly important for rift-type structures which are generally considered to be responses to asthenospheric dynamics and lithospheric thinning.

Satellite magnetic anomaly data from POGO and Magsat have been globally correlated with tectonic structures (Frey, 1979), including the global distribution of rifts by Burke et al. (1978). These average anomaly maps and reduced to pole representations are generalized, yet show intriguing associations with major active and failed continental rifts (Frey, 1981; Von Frese, 1981; Longacre, 1981). Prominent magnetic anomalies are located near the NW portion of the Shatsky aulacogen, the Mississippi Embayment, and the Moma-Zyryanka structure. All of these anomalies are positive, as are the weaker features near the Benue Trough and Amazon River Valley. Major rifts lying along magnetic anomaly contours, often on the gradient between flanking highs and lows, include the Mid-Continent Gravity High, the Pachelma and Moscowian Rifts, and several large structures in the Siberian Platform. These latter represent good examples of the association of satellite magnetic anomalies with tectonic provinces and structure. A major positive anomaly overlies the Siberian Platform, centered near the Anabar Shield. Further east a major negative anomaly overlies the younger Phanerozoic orogenic zone which is separated from the Siberian Platform by the Verkoyansk foldbelt. A major sediment-filled trough mapped by Burke et al. (1979) as a Proterozoic rift lies west of this boundary, east of the Anabar Shield. The magnetic low associated with the orogenic zone deflects northwestward to include this large structure, interrupting an otherwise good correlation between tectonic province and satellite anomaly structure.

More detail is available when selected regional data are studied. As part of our compilation of geologic and geophysical data for a Comparative Atlas of Continental Rifts, we are generating equivalent source, reduced to pole magnetic anomaly maps from POGO and Magsat data, and relating the details of these maps to other geophysical, geologic and geochemical properties. Similar studies have been done for individual anomalies or regions, such as the Mississippi Embayment (Von Frese et al., 1981), the Amazon aulacogen (Longacre, 1981), the Oslo Rift (Ruder and Frey, 1981), Lake Baikal Rift (Frey and Ruder, 1981) as well as continent-wide studies that include major rifts (Mayhew, 1981; Mayhew et al., 1980). We compare the Oslo and Baikal situations here as examples of failed and active rifts.

Figure A shows POGO equivalent source anomaly contours compared with topography and Cenozoic volcanism in the Lake Baikal region. A major negative lies NE of the Lake, along the eastward extension of the faulting and rifting that marks the Baikal Rift Zone. This topographic high has a twin at the western end of the Rift Zone, where rift depressions generally trend N-S as do the POGO negative anomaly contours. For the central, oldest portion of the rift (Kiselev et al., 1978) there is little or no volcanism and a saddle in

Satellite Anomalies and Rifts

Frey, H. and Ruder, M.

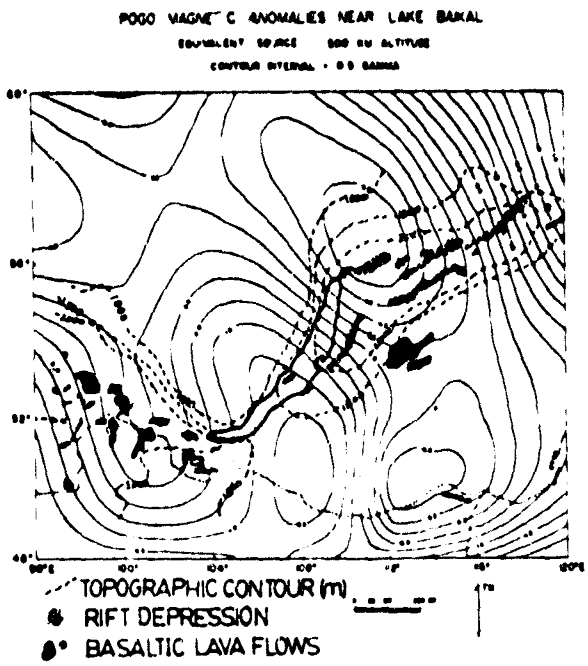

Figure A: POGO equivalent source magnetic anomalies in the Lake Baikal region, compared with topography and Cenozoic volcanism from Logatchev and Florensov (1978). The satellite anomalies are computed at an altitude of 500 km; the contour interval is 0.5 gammas. Note the relative positive anomaly running through south Lake Baikal, where a saddle in the topography occurs and where basaltic volcanism is absent. The strong negative anomalies to the NE and W lie over uplifted regions of more recent rifting (Kiselev et al., 1978).

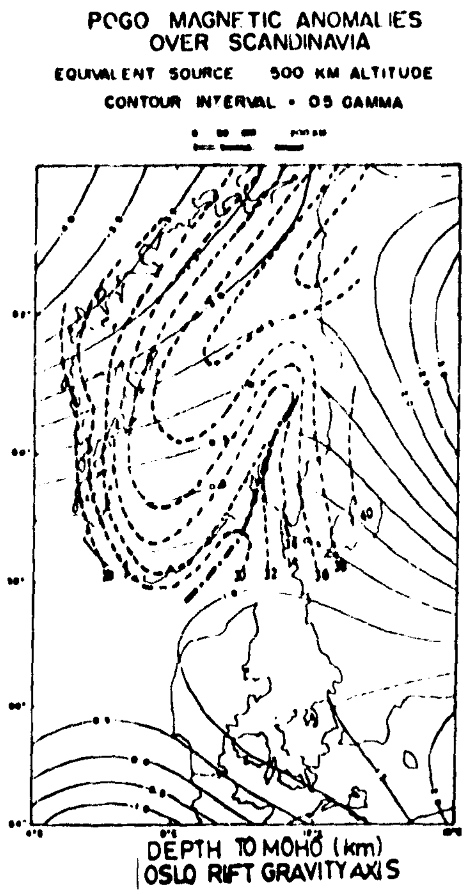

Figure B: POGO equivalent source magnetic anomalies for southern Norway, compared with Moho topography from Ramberg et al. (1977). The crestline of the Moho depths related to the Oslo Rift corresponds well to the bowed contours in the satellite data. This is also the line of Bouguer gravity maxima in this region (Ramberg, 1972).

the topography is evident (Logatchev and Florensov, 1978). This is also the area where a relative positive anomaly appears in the POGO data, running NW through the southern half of the Lake where the Siberian Platform borders the western edge of the Rift. Note that there is no expression of this boundary in the satellite data. There is little additional correlation of the magnetic anomaly pattern with other geophysical data except satellite-derived free-air gravity data from GEM 10C. Heat flow, crustal thickness and seismicity show no obvious relation to the satellite data.

By contrast the satellite magnetic data for southern Norway does show excellent agreement with both depth to Moho and Bouguer gravity. As shown in Figure B a major crest in Moho contours (Ramberg et al., 1977) runs NNE along the Oslo Rift, following almost exactly the bowing in the POGO contours. This is also the line of maxima in the Bouguer gravity signatures for this region, as indicated by the line labeled "Oslo Rift Gravity Axis." As described by Ruder and Frey (1981), a second major structure (the Danish-Polish Depression) runs SE from the Skagerrat and lies beneath the weak positive anomaly that trends along a second line of gravity maxima (Ramberg, 1972) in this region. The close correspondence between geophysical signatures and tectonic structures suggests the magnetic anomaly pattern is likely due to lateral inhomogeneities in the thickness and composition of the lower crust, and we are experimenting with block models to determine the likely range in these parameters that can match the observed data.

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Geologic Resources of the Regions of Late Cenozoic Rifting in Western North America

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There are two regions of Late Cenozoic rifting in western North America. One, the Rio Grande rift, has characteristics typical of most continental rifts, but the other, the Basin and Range province, is a region of broadly distributed extension that reflects ductile flow of the lower crust. Both are well endowed with mineral and energy resources.

These areas have been tectonically active for a long period of time. They have experienced the effects of four distinct plate-tectonic regimes in the Cenozoic era: convergence-related compression; convergence-related intra-arc extension; convergence-related back-arc extension; and post-convergence extension related to development of a major transform fault at the western margin of the North American plate. Long-lived episodic magmatism and high, convective and conductive, heat loss accompanied this history, as did both the emplacement of metals and the shallow thermal maturation of Pre-Cenozoic organic-rich sedimentary rocks.

It was initially believed that a change from calc-alkaline, intermediate to basaltic or bimodal basalt-rhyolite magmatism marked the fundamental change from compressional to extensional states of stress. More recent observations, however, suggest the following relations: (1) magmas emplaced during convergence-related compression were of calc-alkaline andesitic, rhyolitic, and quartz-latitic composition; (2) magmas emplaced during the succeeding period of intra-arc and back-arc spreading at rapid strain rates were of high-silica (locally, peralkaline), rhyolitic composition, accompanied by basaltic andesites, alkali basalts, and locally, tholeiites; and (3) magmas emplaced during the final period of extensional block-faulting that continues today, at reduced extensional strain rates, are of tholeiitic and alkalic basalt composition.

These variations in magma composition with time are reflected in general variations in the composition of some of the emplaced metals. One cannot, however, distinguish between spatial distributions of ore deposits of the convergence-related compressional and extensional regimes, nor between these and the post-convergence extensional regime, on the basis of the general geographic distribution of metal deposits and hydrothermal systems. The spatial distribution of ores of contrasting composition and age (copper, fluorite, and molybdenum) clearly illustrates this point. Metallogenesis in the extended regions appears to be an expression of repeated hydrothermal circulation and recirculation carrying new metals from new magmas upward, recirculation of old metals, and the sweeping of still others from the crustal rocks themselves. These processes continued from an initial period of deviatoric compressional stress into the rift-producing period of deviatoric extensional stress.

REGIONS OF LATE CENOZOIC RIFTING

Eaton, G. P.

Measurable accumulations of syngenetic copper, silver, zinc, molybdenum, vanadium, cadmium, and selenium have been found in local pre-Mesozoic sedimentary rocks in the Great Basin. There is clear evidence that such metals can be, and have been, remobilized, at least locally. Pre-existing structures and major structural flaws apparently served as local zones of continuing or renewed permeability, enhancing the circulation and migration of fluids containing such metals.

Grabens and asymmetric rift structures of Late Cenozoic age served as vessels for the accumulation of both clastic sediments and commercial deposits of a great variety of evaporites. Two cycles of petroleum generation appear to have taken place in the region, one in early Mesozoic time, during the compressional regime, the other, in Late Cenozoic time, during the extensional regime. Production and reserves of hydrocarbons are sparse, however, perhaps owing to long-continued, episodic fracturing and hydrothermal circulation and the absence (or lack of integrity) of an adequate stratigraphic seal.

The geographic extent of both the Basin and Range province and Rio Grande rift appears to have been predetermined in large part during the compressional regime by the integrated sum of successive invasions of the continental crust by magmas of intermediate composition. They weakened the lithosphere and gave rise to a thermally activated, rheologically layered crust. The emplacement of these magmas was, in turn, dictated partly by still older continental structure. Magmatism spanned the full range of the Cenozoic Era, as well as the latter part of the Mesozoic Era. The magmas provided a vehicle for the mass transport of heat to shallow depths, and with them, a driving force for hydrothermal convection in the shallow crust.

PORPHYRY-MO OCCURRENCES IN THE OSLO RIFT-SYSTEM

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Continental rift environments are known to be potential areas for generation of a variety of metal and fluorite deposits. Porphyry deposits are usually related to convergent plate environments, however some porphyry-Mo deposits seem to be related to continental rifting. This study deals with recently recognised porphyry-Mo deposits of the Oslo rift and their occurrence in the rift environment.

The overall structure of the Oslo rift-system can be viewed as reflecting the result of interference between a N-S trending Permian fault system and a regional NE-SW trending, Precambrian shear zone. The zone of N-S trending master faults which follows the Oslo fjord divides the province into two segments which are situated slightly en enchelon, but possess a notably symmetrical distribution of rocks and structures. The axial zone is occupied by Cambro-Silurian sediments which have been intruded by major granite bodies. Extensive lava-plateaus occur on each side of this axial zone. The subsequent zones contain the most prominent cauldrons of the province. Further away from the axis follows a zone of batholitic intrusions, emplaced as composite diapirs and plutonic ring complexes. Finally, at the lateral borders of the province, occur hornfelsed sediments and volcanics in narrow bands, possibly preserved as the result of marginal tectonics associated with batholith emplacement.

Mo-mineralizations are almost exclusively associated with subalkaline granitic members of the strongly diversified rock--assemblage of the province and are found in 1) the granites of the axial zone; 2) subvolcanic acid rocks of the cauldron zone and 3) plutonic ring structures of the batholith zone. Porphyry type Mo-deposits, however, are strictly associated with subvolcanic environments in the latter two zones. Two important prospects will be discussed: the Glitrevann occurrence in the cauldron zone of the southern segment and the Nordli deposit in the batholith zone of the northern segment.

The porphyry-Mo mineralization associated with the Glitrevann cauldron is related to the granitic stock which occupies the central part of the cauldron. The evolution of the cauldron can be divided into three stages: 1) Caldera-forming eruptions of lapilli-ash flows of trachytic and rhyolitic compositions together with extrusion of trachytic and basaltic lavas. Along the margin of the caldera, mega-meso breccias were formed, probably as a result of caldera collapse; 2) intrusion of a syenitic ring dike which postdates the extrusivas; 3) intrusion of the central granitic stock, post-dating the ring dike. The central stock is a multiple intrusion composed of medium to coarse grained granite, porphyritic granite and aplogranite. The aplogranite occurs in a subcircular zone in the central part of the stock, and is the youngest of the intrusive phases. Mo-mineralization is spatially related to the aplogranite and occurs especially along the outer rims of the subcirkular zone. At

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Bordvika, where a massive, rhyolitic ignimbrite covers the aplogranite, there is a stockwork of Mo-veins enveloped by sericitic alterations. The sericitic altered area has a diameter of approximately 1 km, but the Mo-stockwork only occurs in the central part of this alteration, whereas disseminated molybdenite rosettes are found in the outer part of the sericite zone. The sericite alteration is surrounded by a poorly defined argillitic alteration zone which is associated with a heavy manganese staining; this marganese staining is locally so extensive that it was mined at the beginning of this century.

The Nordli porphyry-Mo deposit is associated with a composite, central-intrusive stock of alkali-granitic composition occurring within batholith syenites and associated effusives of the northern Oslo province. It is related to the formation of regional ring structures, and forms the youngest regional magmatic event in the area. The central stock formed by multiple injection and consists of a succession of compositionally similar effusive and intrusive materials which have highly variable textures. In the central part masses of polylithological, unsorted and partly unconsolidated ash-lapilli tuffs are entirely enclosed and invaded by younger quartz-feldspar porphyries which show elements of both intrusive and extrusive characters. The quartz-feldspar porphyries are subsequently intruded, and partly assimilated by a medium-grained, central alkali-granite. This alkali-granite also occupies an external circular fracture forming a ring complex of about 15 km diameter, and with the central stock slightly off-center to the north. The subsequent quartz-eye porphyritic granite and microgranophyre, separated by subvolcanic brecciation, form essential elements of the closing igneous events in the central stock. The Mo-mineralization is associated with extensive hydrothermal alterations which postdate the entire rock suite. They are primarily hosted within the quartz-eye granite and the microgranophyre and are associated with a well-developed mineralized stockwork of intense sericite alteration. The alteration is reflected by a circular negative magnetic anomaly which cross-cuts several lithological boundaries and are rimmed by concentric marginal pyrite- and manganese-haloes.

The porphyry-Mo mineralizations in the Glitrevann and Nordli prospects have several evolutionary features in common. 1) An initial period of effusive eruptions of pyroclastic flows followed by 2) the formation of large-scale ring structures, partly accompanied by peripheral syenite-granite intrusions, 3) the emplacement of a multiple intrusive, central stock of subalkaline granite composition which is terminated by 4) the injection of a highly differentiated aplogranite/granophyre, closely associated with the development of the mineralizing hydrothermal porphyry system.

The distribution of Mo-mineralizations in the Oslo province shows that these are closely related to granitic rock members, specifically of subalkaline affinity. However, the formation of proper porphyry-type Mo-deposits additionally requires the position in a subvolcanic environment such as found in many of the cauldrons and ring complexes.

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The discovery of several significant porphyry-Mo type mineralizations in the Permian Oslo rift draws attention to the possible economic potential of rift structures elsewhere.

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METALLIFEROUS RESOURCES ASSOCIATED WITH RIFTING : THE PROTO NORTH ATLANTIC EXAMPLE (360-280 Ma)

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The genesis of two types of mineral deposit related to a single period of rifting in northwestern Europe are considered: (i) economically important sediment-hosted exhalative Pb + Zn +$BaSO_4$ ± Cu deposits generated early in rift development, such as the 10^7 tonnes of Zn + Pb metal at Navan, Ireland; (ii) sparse minor vein silver deposits spatially associated with the 295 Ma quartz dolerite dykes that mark the end of continental rifting. The model for type (i) deposits discussed below is also applicable to the generation of other major exhalative Pb + Zn deposits, e.g. Mt Isa, McArthur River and Broken Hill (Australia) and Sullivan (Canada)(Russell *et al.* 1981).

In the earliest Carboniferous much of Ireland and parts of Britain were subsiding slowly and uniformly and were covered by a shallow highly saline sea. The sudden onset of rifting at 360 Ma was accompanied by alkali basaltic volcanism and spatially unrelated exhalative sulphide mineralization (Table 1). That the major base-metal deposits in Ireland are exhalative is demonstrated by (i) the discovery of hydrothermal pyrite chimneys at Silvermines (Fig. 1) similar to those at 21°N on the EPR (Carter *et al.* 1981), (ii) primary manganese aureoles extending radially 7km from Tynagh (Russell 1975) and (iii) sulphide clasts in submarine debris flows in all the major Irish deposits. The deposits (Fig. 3) formed contemporaneously in local basins on the sea floor adjacent to active portions of major fault zones. Temperatures of the hydrothermal solutions reached at least 265°C (Samson 1980) and vertical metal zonations imply a rise in temperature with time during mineralization at Tynagh and Silvermines. These features are explained by deriving the ore solutions from convective circulation of modified saline seawater in the thick (≤15km), hot, 'fertile' L. Palaeozoic geosynclinal prism (stablized at 400 Ma) constituting the upper crust beneath Ireland (Russell 1978, Russell *et al.* 1981).

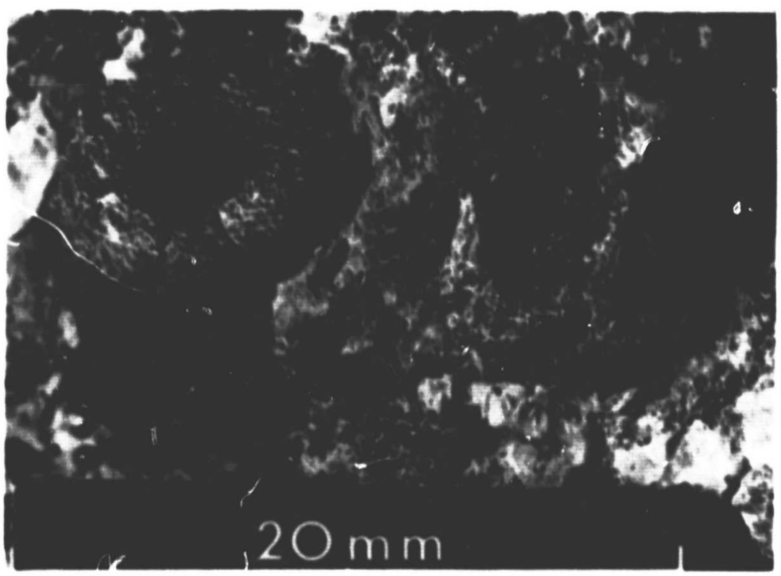

Fig. 1 Pyrite chimneys from Silvermines

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TABLE 1 Correlation between tectonic, magmatic and metallogenic events in Ireland, north Britain and Norway.

AGE Ma	TECTONICS	MAGMATISM	INFERRED STRESS REGIME	INFERRED REGIONAL TECTONICS	MINERALIZATION
270	Fault bounded	------------	Unstable young	~150km of ocean	
280	NRS basins UK	Oslo Graben	continental	floor spreading	Vein Ag+minor
290	Uplift in UK	syenites Dykes	margin NS tension UK	Final separation	fluorite
300		Intermitt-	Stretching:-	~100km spreading	
310	Oslo Graben	ent alkali	minimum stress		
320	Coal Basins	basalts		Culmination of	
330		-----------	trajectories	continental	
340		Main period	reorienting	rifting UK	
350		of alkali	between E-W	General basin	
360	Starved basins	basalt volcanism	and NW-SE	and rift development	Exhalative Pb+Zn+Ba±Cu

The hydrothermal convection cells were initiated during early rifting, were driven by heat in the rock column, and the updraughts located at fracture intersections or particularly permeable sections of fault zones. Some of the operative fractures had trends similar to those that now define the continental margin to the northwest (Russell 1968). As a result of continued extensional strain and cooling of the upper crust the brittle-to-ductile transition zone was depressed and the circulation penetrated to greater depths with time (Fig. 2). This theory is consistent with oxygen, hydrogen (Samson 1980) and lead isotope studies (Boast *et al.* 1981).

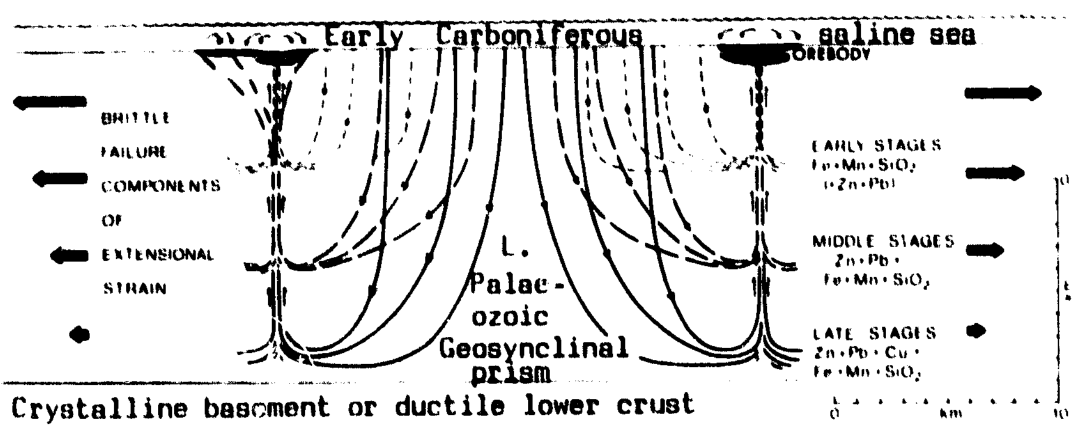

Fig. 2 Generalized model for downward-penetrating convection cells feeding Irish type deposits; metals are leached from the Lower Palaeozoic geosynclinal metasediments (after Russell 1978, Russell et al. 1981).

We may expect the temperature of the solutions to have risen with time as they encountered and leached hotter rocks at depth so accounting for the late addition of the less soluble metals, Cu and As. Although the locations of the ore deposits are structurally controlled the dimensions of fully developed convection cells dictate a minimum spacing between major deposits of about 20km.

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After the cells reached their maximum dimensions the convecting solutions entered the cooling stage so giving rise to the hanging wall aureoles. On occasion, where the flow paths had become oxidized, a later copper phase may have followed (e.g. minor chalcopyrite in late dolomite at Tynagh). There were no further significant mineralizing events although submarine rifting continued for ~35My and subaerial rifting for another ~15My (Table 1). It apparently required heat from the intrusion of quartz dolerite dykes at 295 Ma to re-engender hydrothermal convection, albeit of relatively low temperature. These fluids were only capable of dissolving Ag± minor Pb,Zn,Cu, Co,Ni,As and Ba from underlying fertile sediments and metasediments to produce the vein deposits at Kongsberg,Norway, and the Ochils, Strontian, Hilderstone,in Scotland.

We have argued that these dykes, as well as coeval flows in the Oslo Graben, marked the time of final lithosphere separation in the 'pre anomaly 24' Proto North Atlantic (Russell and Smythe 1978) and that they are the result of a stress system of radial minimum principal stress focussed on the Faeroes area. In this region there is a mismatch in the otherwise 'parallel' sided pre anomaly 24 ocean which suggests significant crustal stretching while ~100km of ocean floor spreading was taking place to the northeast and southwest (Fig. 3 and Table 1). This hypothesis has led us to predict successfully a major extension of this dyke swarm into the North Sea and reveal its arcuate trend (Fig. 3). A further ~150km of ocean floor spreading followed final rupturing in this region of Archaean crust. The model is consistent with right-lateral motions in northern Spain between ~315 and 300 Ma (Heward and Reading 1980), and 150km of right-lateral movement along the Proto Bay of Biscay Fault between 295 and 250 Ma (Arthaud and Matte 1977).

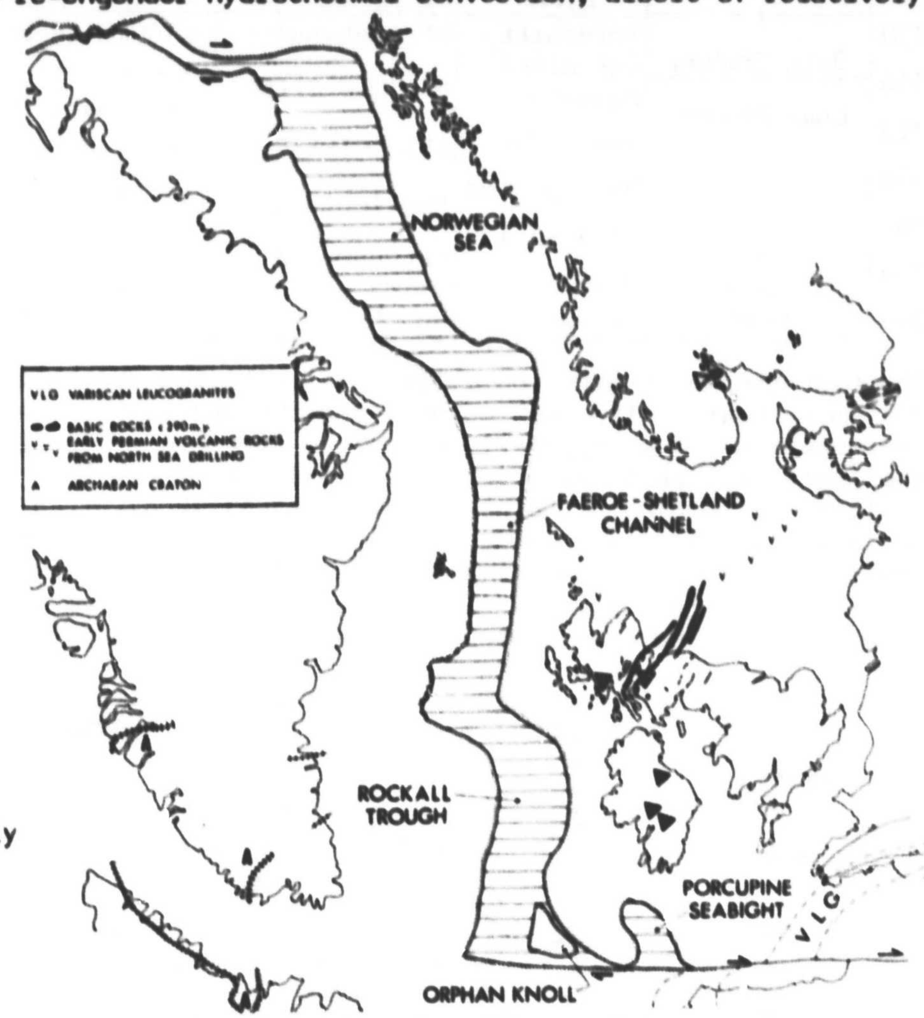

Fig. 3: Pre-Tertiary, U.Carboniferous (?) Proto North Atlantic Ocean. ▲ *major 360 Ma exhalative base-metal deposits,Silvermines,Tynagh,Navan (south to north),* △ *minor silver deposits (~290 Ma ?).*

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Some fluorite mineralization was associated with the alkali intrusives in the Oslo Graben (~280 Ma) but the major fluorite mineralization of the Pennines of England, originally widely assumed to be of early Permian age, may be as young as upper Triassic (205±7 Ma) according to the Rb/Sr dating of fluid inclusions by Shepherd and Darbyshire (1981). If so it may be related to the initiation of a second major rifting phase focussed further south in what was to become the Central Atlantic.

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ACCUMULATION OF FOSSIL FUELS AND MINERALS IN ACTIVE AND ANCIENT RIFTS

by Eleanora Iberall Robbins, U.S. Geological Survey, Reston, VA 22092

Organic fuels and minerals accumulate in sediments and rocks of both active and ancient rifts. For example, petroleum, oil shale, lignite, and bituminous coal are exploited in the Rhinegraben rift in Germany. In the Jordan-Arava rift, widespread deposits of peat and lignite underlie Lake Hula in Israel, and evaporite minerals such as potash are extracted from brines in the Dead Sea. Coal, phosphate, and nitrogen-rich black-shale fertilizer were mined in North Carolina in the lakebeds of the Triassic Cumnock Formation in the Deep River basin of the Newark rift system (see Robbins, this conference). Much of the current prospecting in the Newark rift system is for economic deposits of uranium. Lead and zinc are mined in Australia from the Urquhart shale of Proterozoic age in the Leichhardt River rift. Copper is mined in Michigan and Wisconsin in the lakebeds of the Nonesuch Shale of Proterozoic age in the Midcontinent rift.

A study of active and ancient rift systems suggests that the accumulation of organic and mineral resources is related to the interactions of processes that form rift valleys with those that take place in rift lakes. The interacting processes include tectonic, thermal, climatic, hydrologic, sedimentological, limnological, chemical, and biological factors. The most important interactions revolve around faulting, which causes uplift and shatters rocks. Weathering and erosion result in the release of nutrients and metallic ions into the watersheds of tectonic lakes. Nutrients carried downstream may increase the productivity of organisms that are precursors of fossil fuels. Metallic ions are known to precipitate depending on redox conditions that are enhanced by high organic production.

Rift systems are characterized by linear rupture in the crust and by high heat flow. Periodic earthquake activity can result in uplifted highlands and horsts and down-dropped grabens and tilted fault blocks. Hot springs and geysers may rise along faults, and volcanic activity may lead to emplacement of intrusive and extrusive igneous rocks. The bordering highlands are eroded at increased rates, and the valleys are filled by coarse- to fine-grained clastic continental deposits where dry, and by lacustrine deposits where impounded freshwater lakes have formed. Tremors accompanying faulting are known to produce landslides and slumping, and to agitate bottom sediments that may release nutrient-bearing pore waters into lakes.

Precipitation may be modified near rift systems because newly formed highlands and large volcanic cones alter weather and drainage patterns. High-gradient streams that carry large suspended and solute loads from the faulted and weathered highland rocks often deposit the sediment as deltas in the rift-valley lakes. Near-surface ground water is driven under higher hydraulic head from the higher valley shoulders and can leach more of the soluble ions from the rocks. Deep-seated ground water may be heated to form circulating convection cells, also effective in leaching rocks. As a result, the surface- and ground-water regimes of rift valley lakes differ from those of other freshwater lakes, notably

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in their relatively high content of total dissolved solids (TDS) and also in their relatively high pH values. Of 31 modern rift valley lakes, 29 are alkaline. Those lakes having an input of warm or hot waters can become thermally stratified and eventually also density (salinity) stratified. The bottoms of strongly stratified lakes tend to be anoxic environments favorable to growth of anaerobic bacteria and an accompanying generation of both methane and H_2S. The salinity of rift-valley lakes in desert regions increases where evaporation exceeds inflow, attaining the highest levels in the Dead Sea.

Much chemical activity results from these geologic events, starting with the weathering of rocks in the highlands. Should the highlands consist of crystalline and volcanic rocks of pre-Mesozoic age, as they did along the Newark rift system, such minerals as allanite, uraninite, sphalerite, zinc-bearing magnetite, zircon, and chalcopyrite would be available to be leached. Concentrations of uranium, copper, and zinc have been identified in "black" profundal shale, profundal and littoral siltstone, and gray littoral sandstone of former lakes in the Newark rift system. Also released in the weathering process are trace elements needed for the growth of organisms including phosphorous from such minerals as apatite.

Autochthonous organic matter also accumulates in rift valley lakes. In those lakes where waters contain high concentrations of dissolved solids, biological activity is correspondingly high. Biological productivity is expressed as biomass, production of organic carbon, or the length of the ensuing food chains. A lake containing a large population of photosynthetic algae can support a large population of predacious zooplankton. A large zooplankton community can support a large fish population. Large fishing industries in such lakes as Lakes Baikal and Malawi attest to high biological productivity in modern rift valley lakes. A large fish population can support a large carnivore population; in the last census of Lake Rudolf, more than 12,000 adult crocodiles were counted. The death and decay of such a large biomass can result in the water near the bottom and the lake sediment becoming anoxic, thereby preserving the precursors of petroleum, oil shale, and phosphate.

The community of plants and organisms probably determines the chemical composition of the resulting organic deposits. Certain plants, such as Sphagnum which grows around Lake Lungwe (between Lakes Tanganyika and Kivu), cause acidic conditions in swamps by a process known as ion exchange. Organic tissues are known to be preserved best under acid conditions. The green alga Botryococcus, which proliferates under alkaline, saline conditions of lakes where rainfall is limited, stores petroleum-precursor lipids and hydrocarbons. The high alkalinity and salinity of the Dead Sea are ideal for the growth of the dinoflagellate Dunaliella, another lipid-storing alga. However, such conditions are unfavorable for the accumulation of petroleum, because alkaline solutions can saponify and, therefore, destroy the petroleum-precursor lipids. Carbohydrate-storing, blue green algae may be precursors of oil shale. The fecal pellets of algae-eating zooplankton are a suggested major source of phosphate in the lacustrine shale of the Deep River basin in the Newark rift system.

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The four most common metals found in rift-valley deposits are copper, zinc, lead, and uranium. These metals are known to precipitate in anoxic water, or at contacts between anoxic and oxygenated waters in lakes today. Aerobic-anaerobic interfaces are found 1) at the sediment/water interface along the edges where lakes are anoxic down in the water column, or at depth where lake bottoms are oxygenated; and 2) along the margins of lakes where anoxic lake water seeps into and intersects the ground water. In addition to precipitating at these interfaces, metals can also precipitate where the products of decaying vegetation in swamps produce humic acids that are washed into lake waters. Uranium and copper are known to complex with humic acids at such interfaces.

Other sulfides are formed directly where metallic ions encounter H_2S generated by anaerobic bacteria. Zinc carried by hot springs precipitates as minute spheres of sphalerite around bubbles rising in the H_2S-rich bottom of Lake Kivu. Similar blebs of a copper-colored mineral are found in the Nonesuch Shale in the Midcontinent rift.

During the active stages of rift development, sediments that contain organic fuel precursors and minerals accumulate at different places. Peat is deposited primarily in swamps associated with river deltas, but also along lake embayments and inflowing rivers. Uranium and copper precipitate at aerobic-anaerobic interfaces along lake margins or within lakes. The petroleum and phosphate precursors are deposited in anoxic parts of lakes. Copper, zinc, and lead sulfides also may be deposited in anoxic waters, principally near source hot springs.

The organic accumulation can be tranformed into fossil fuel during the active stage of rifting. The plants of the swamps can be buried rapidly, and the rank can change quite drastically. Lake Hula contains 111 m of peat near the surface and lignite at depth. The widespread deposits of lignite (Braunkohle) in the Rhinegraben have been tranformed to subbituminous and bituminous coal (Steinkohle) at depths of 1,600 m around hot spots such as Landau. Bituminous coal, semi-anthracite, and anthracite are found today at the surface in Triassic basins of the Newark rift system.

Petroleum-precursor organisms in buried lakebeds may be transformed by burial, pressure, and temperature in time to liquids and gases, and finally just to gases in a process known as maturation. Temperatures in the range of 66-132°C, depths greater than 1,200 m, and some minimal amount of time, usually cited as 10,000 years, are the requirements generally cited for petroleum generation. These conditions are met around the Landau hot spot in the Rhinegraben, where lacustrine source beds of Oligocene age produce petroleum at depths between 1,900-2,200 m.

Petroleum accumulation requires several other factors besides organic-matter-rich source beds. The potential of petroleum accumulation in an area is directly related to the presence of porous and permeable reservoir rocks, structural and/or stratigraphic traps, and seals. Deltaic deposits in large lakes can be large accumulations of relatively coarse-grained, permeable clastic materials. Subsidence, landslides, and slumping may also provide porous materials. Sealing materials on river deltas are commonly clays that disperse out from flood waters in the upper delta plains. Clays also flocculate where freshwater rivers enter alkaline lakes. Faults form traps when nonporous and porous beds are brought into contact. Faults may also eventually seal because they can serve as conduits for charged ground water.

There is probably also some post-depositional rearrangment of metals due to increased temperatures and heated ground-water flow through porous and permeable layers at depth. In many coal beds, megascopic pyrite is found in cleats (joints). Much of this pyrite may have been initially deposited as point-particle framboids from bacterial metabolism of organic tissues. Some of the round, octahedral, and pyritohedral holes in the organic matter forming coal in the Dan River-Danville basin in the Newark rift system suggest loci from which pyrite was leached.

The interaction of physical, chemical, and biological processes in rift valleys therefore has produced accumulations of minerals and fossil fuels. A better understanding of the depositional and post-depositional processes will help in planning exploration programs when looking for mineral and organic fuel deposits in ancient rift systems.

MAGMA COMPOSITION VARIATIONS ALONG HOTSPOT TRACES: PALEOLITHOSPHERIC THICKNESS AND COMPOSITIONS OF THE LVZ

By R. F. Wendlandt and C. Podpora

Gass et al. (1978), using thermal conduction models to determine the thermal disturbance in lithosphere over a hotspot, noted that lithosphere thinning could be accomplished by deep, strong thermal peturbations, affecting the lithosphere over, perhaps, a radius of 500 km, or, thinning could be accomplished by an ascending thermal anomaly of a lesser magnitude affecting a much smaller radius. They noted that only slight motion of lithosphere relative to the hotspot (>2 cm/year) is needed to suppress asthenospheric upwelling and volcanic activity. In a previous abstract (Wendlandt, this volume), volcanism associated with Mesozoic and Cenozoic African rifting, occurring during eras of minimal plate-hotspot relative motion, was used to constrain physical aspects of lithothermal systems. In this abstract, systematic variations in intraplate volcanism in Brazil and S. Africa, arising from the motion of lithosphere over a hotspot, will be used to constrain physical aspects of the lithosphere and the chemistry of partial melts in the low-velocity zone (LVZ).

Hotspot traces have been suggested previously for S. Africa (Rhodes, 1971; Duncan et al., 1978; Duncan, 1981) and Brazil (Crough et al., 1980). Igneous provinces associated with the passage of S. Africa over a hotspot are summarized in Fig. 1A: the opening of the S. Atlantic about 125 m.y. b.p. and the motion of Africa in a northeasterly direction (at a rate of approximately 2 cm/yr; Minster et al., 1974) defines a trail of predominantly Cretaceous igneous occurrences extending form the vicinity of the Karroo flood volcanics to the current hotspot location northeast of Bouvet Island (Meteor Hotspot; Crough, 1979). Crough et al. (1980) also identify a Jurassic volcanic trace along approximately the same trend, terminating at Bouvet. In an analogous fashion, the S. American plate has moved in a westerly and northwesterly direction at 2 cm/yr (Crough et al., 1980; Minster et al., 1974) and a trace of igneous events extends from south-central Brazil to the present hotspot location near Trinidade (Fig. 2A).

A relation between the Karroo flood eruptives and the S. African hotspot is implicit. The location of the hotspot prior to the break from S. America can be extrapolated to the region immediately west of Lebombo and north of Lesotho. Briden and Gass (1974) have proposed that the African plate was approximately motionless at this time which would enable lithosphere thinning and eruption of flood volcanics, and Fairhead and Reeves (1977) have contoured a thinned lithosphere in this region of southern Africa.

With the initiation of plate motion igneous petrogenesis is controlled by the depth of the lithosphere-asthenosphere interface since it is unlikely that conductive heating of the lithosphere will result in partial melting (Gass et al., 1978). The chemistries of the eruptives, therefore, are an indication of the thickness of the lithosphere at the time of melt segregation and of the compositions of melts in the LVZ. In S. Africa, a systematic transition toward the coastline (summarized in Fig. 1B) from diamond-bearing kimberlites to diamond-absent kimberlites to carbonatites and strongly alkalic silicate melts to trachytes along and just seaward of the coastline, is indicative of decreasing depth of partial melting. The basis for estimating depths of melting and a discussion of estimate validities are presented in the paper by Wendlandt (this volume).

The Brazil volcanic trace is more complicated. The general trend is one of decreasing depth of magma origin toward the coast with time (Fig. 2B);

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however, perturbations in the trend are observed. The oldest (c. 100-80 m.y. b.p.) and furthest from the coast likely represents interaction of the hotspot with an existing section of attenuated lithosphere. Lithosphere attenuation associated with this older event (eruption of the Parana basalts) registers as a temporary decrease in depth of magma origin. The second perturbation is manifest as a scattering of the distributions of occurrences (Fig. 1A) and depths of origins affecting occurrences dated between about 80-65 m.y. This interval coincides, however, with the cessation of the motion of the S. American plate, eventually reinitiating with a new pole of rotation (Dietz and Holden, 1970; Nairn and Stehli, 1973) The scatter of rock ages and distributions in Brazil at this time, may reflect lithosphere thinning, and concommitant asthenospheric upwelling, associated with the lack of plate motion. The apparent inflection in the path of the lithosphere over the hotspot, represented by the E-W trend of alkaline occurrences near the coast, may not indicate a real change of plate motion; igneous distributions may be surfacially controlled by existing structures, including the landward extensions of oceanic fracture zones (Sykes, 1978). Lithosphere thicknesses are shown by this analysis to decrease gradually across Atlantic-type continental margins.

The composition of melt in the LVZ has been variously debated. Kushiro et al. (1968), Lambert and Wyllie (1968), and Green (1970) suggested that the LVZ resulted from amphibole instability; Green (1970) proposed that melt compositions would be nephelinitic or melilititic. Basu and Murthy (1977) suggested that the composition of partial melt might be kaersutitic. Olafsson (1980) has demonstrated that incipient melting of an amphibole-bearing lherzolite is likely to be nephelinitic. Wyllie and Huang (1975) and Eggler (1976) have also suggested that the LVZ may result from partial melting of a carbonated peridotite at pressures in excess of 25-30 kbar; the anticipated melt composition being strongly carbonate-normative and perhaps carbonatitic. If the results of analyzing magmas from hotspot traces are applicable to this debate (i.e. it is assumed these melts are largely derived from the zone of asthenosphere-lithosphere decoupling), then melt compositions in the LVZ are variable, ranging from kimberlitic under thickest lithospheres through carbonatitic and nephelinitic to phonolitic and perhaps trachytic compositions under thin lithospheres. The relation between kimberlites and hotspots has been noted previously by Crough et al. (1980), and Anderson (1975; 1981) and the control of lithosphere thickness on kimberlite distributions has been suggested by Eggler and Wendlandt (1979). This study suggests that diamond-bearing kimberlites require lithosphere thicknesses in excess of approximately 170 km (Kennedy and Kennedy, 1976) at the time of melt genesis.

The results of this analysis support the model of Gass et al. (1978) in which even small rates of plate motion over a hotspot were believed sufficient to suppress lithosphere thinning and, thus, the possibility of eventual rifting. Where the circumstances are appropriate, however, the correct combination of slow plate motion, thin lithosphere, and a strong thermal perturbation may result in rifting and rift jumping (e.g., Hey et al., 1977; Fitton, 1980; Wood, this volume).

MAGMA COMPOSITION VARIATIONS

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A

B

Fig. 1 (A) Igneous provinces possibly related to the passage of S. Africa over hotspots. (B) Section across the continental margin showing depths of origin of magmas. Occurrences are projected short distances onto the line of the section shown in (A). Temporal and spatial relations of volcanics are from Nicholaysen et al. (1962), McDougall (1963), Gough et al. (1964), Manton (1968), Cox (1970, 1972), Dawson (1970), Fitch and Miller (1971), Dingle and Gentle (1972), Allsopp and Barrett (1975), Davis et al. (1976), Allsopp and Kramers (1977), Davis (1977a,b), and Duncan et al. (1978). The limit of the Transvaal Craton is from Cahen and Snelling (1966).

MAGMA COMPOSITION VARIATIONS

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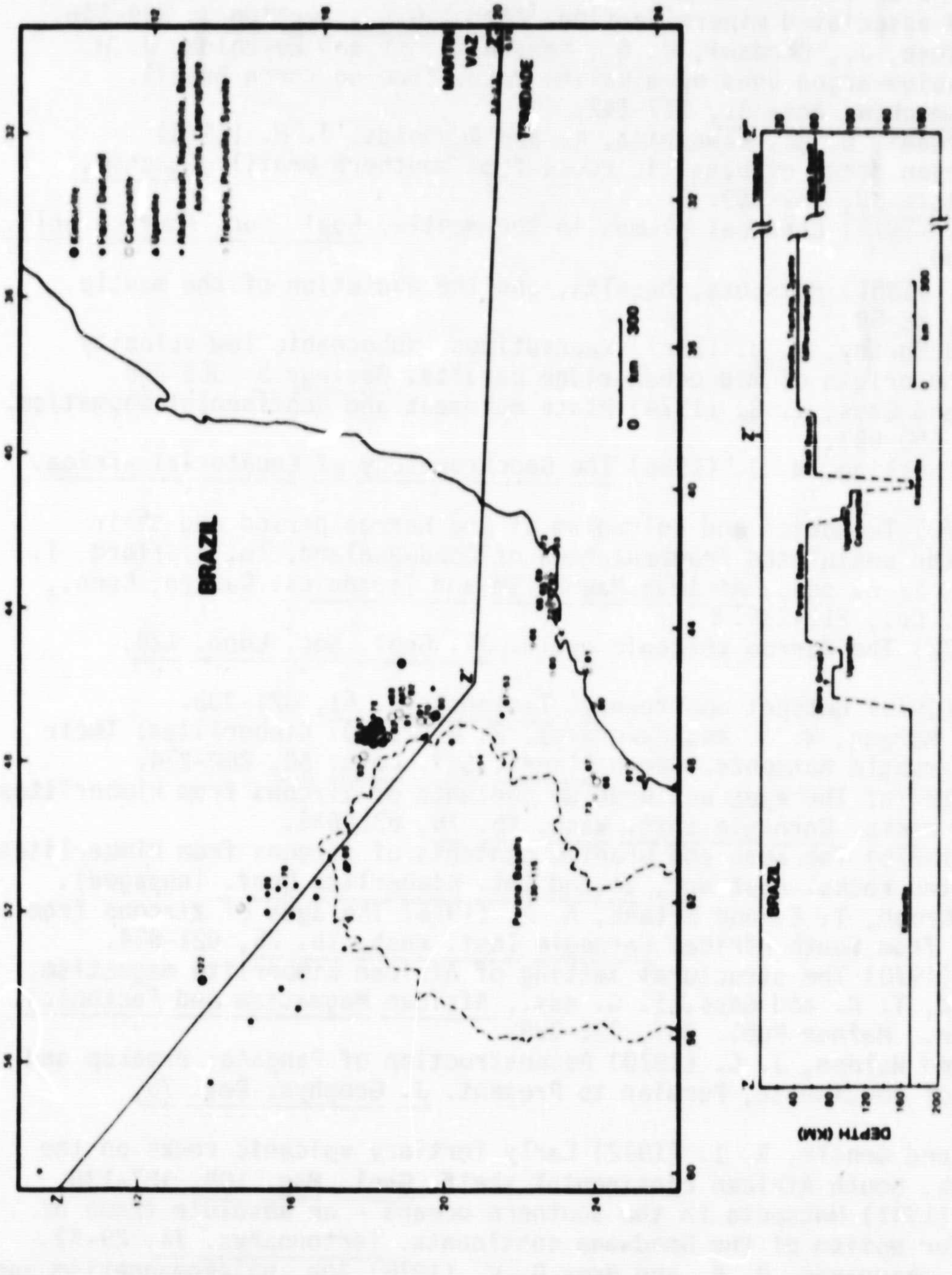

Fig. 2 (A) Igneous provinces possibly related to the passage of Brazil over a hotspot. (B) Section across the continental margin showing depths of origin of magmas. Occurrences are projected short distances onto the line of the section shown in (A). Temporal and spatial relations of occurrences are from Amaral et al. (1966), Amaral et al. (1967), Almeida (1972), Davis (1977a,b), Herz (1977), Svisero et al. (1979), Crough et al. (1980), and Ulbrich and Gomes (1981).

MAGMA COMPOSITION VARIATIONS

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THERMAL UPLIFT AND SUBSIDENCE AT CONTINENTAL RIFTS DUE TO HORIZONTAL EXTENSION OF THE LITHOSPHERE

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The formation of grabens in continental crust is generally associated with tensile faulting, crustal thinning, anomalously warm mantle below the rift zone and "updoming" of the outer flanks of the rift zone. These features are also common to the initial stages of sedimentary basin formation and continental rupture (creating new sea floor). This similarity suggests that similar tectonic processes, but of varying duration, may be responsible for all three phenomena. There is now mounting evidence that the latter two result from localized horizontal extension of the lithosphere followed by cooling of hot mantle material which upwells passively during the extensional phase.

This study addresses the question of whether the tectonics associated with graben formation might also be induced passively by horizontal extension of the lithosphere and crust. Clearly horizontal extension can account for tensile faulting, crustal thinning and a narrow zone of anomalous mantle immediately below the graben. A key question however is: can such a process account for the regional updoming which is invariably observed about young grabens and a broad zone of anomalous mantle below the rift?

The physical model of the rift process employed here is the same as that suggested by McKenzie (1978) for the initial stage of basin formation. A narrow block of lithosphere and overlying continental crust is suddenly stretched by a factor β, causing a thinning of both lithosphere and crust by the same factor β. Extension is accommodated in the upper brittle portion of the continental crust by listric normal faults and at greater depths by ductile flow.

Within the zone of extension two competing effects act to determine the sense and magnitude of initial epeirogenic movements which are required to maintain isostatic equilibrium. These are (a) the thinning of the low density crust (which alone would result in initial subsidence) and (b) the upwelling of hot asthenospheric material below the rift in order to compensate for the horizontal outflow of material in the stretching zone (which alone would result in initial uplift of the rift zone). Where the crust is very thin, as on the ocean floor, the second effect dominates and initial uplift such as that occurring at mid-ocean ridges can be expected. Where the crust is thick, however, either subsidence or uplift may initially occur depending on which effect is dominant.

Continental rifts differ from mid-ocean ridges in that each rift results from a discrete stretching event rather than from a continuous process. Thus ultimately, regardless of whether the initial response to extension is uplift or subsidence, diffusive cooling of the anomalous mantle ensures that subsidence and graben formation occur. Both the initial and final responses to extension are confined to the region in which crustal thinning occurs. However transient effects (lasting 60 Ma or more) may occur on the outer flanks of grabens due to horizontal diffusion of heat outwards from the anomalously hot material below the rift. The horizontal heat flux is most

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significant when the transition from rifted to non-rifted crust is very sharp, since horizontal temperature gradients are then maximized. It results in a temporary warming of the lithosphere and crust adjacent to the rift zone and an increased rate of cooling of the anomalous mantle below the rift zone. Isostatic adjustment during this transient phase results in uplifted outer flanks of the graben structure and a relatively rapid thermal subsidence of the graben floor. The topography of the graben floor becomes convex upwards due to a more rapid cooling of the edges than the centre of the rift zone.

Time-dependent two-dimensional numerical solutions are obtained for the temperature field in the crust and lithosphere in a broad region centred on the rift axis. Temperature is assumed to be invariant along the strike of the rift and symmetric about its axial plane. Thus numerical solutions need only be obtained in terms of the vertical coordinate, z, horizontal coordinate measured normal to the rift axis, x, and time t. Before rifting, the temperature field is assumed to vary linearly with depth from $T = 0°C$ at the upper surface to $T = T_1$ (a constant) at the base of the lithosphere. Immediately after rifting (i.e. stretching by a factor β) a rift zone of width $2W$ exists centred on the ridge axis. For $x>W$ the (unstretched) thicknesses of the crust and lithosphere are t_c and a, respectively. Hence for $x<W$ crustal thickness becomes t_c/β and the depth to the base of the lithosphere a/β. The stretching is assumed to have occurred instantaneously so that temperatures are unaffected for $x>W$. However for $x<W$, temperature now varies linearly from $T = 0°C$ at the upper surface to $T = T_1$ at a depth a/β and is assumed to remain constant at $T = T_1$, the temperature of the asthenosphere, from the depth a/β down to the depth a. This initial temperature field is allowed to evolve diffusively with time. At any time the temperature solution can be integrated to infer the isostatically adjusted topography across the rift zone and environs.

Both the amount and duration of uplift adjacent to the rift zone depend strongly on the depth, a, of the lithosphere (and the stretching factor β). When $a = 125$ km and $\beta = 2$, a maximum thermal uplift of 0.6 km occurs at $t = 7$ Ma and diminishes to 0.4 km at $t = 30$ Ma and 0.2 km at $t = 60$ Ma. The width of the uplifted zone is approximately 50 km wide. However for $a = 250$ km and $\beta = 2$ maximum uplift is 1.3 km at 7 Ma diminishing to 0.9km at $t = 60$ Ma and 0.6 km at $t = 100$ Ma. The width of the uplifted zone for this (extreme) case is $\sim$100 km.

Both the initial and total subsidence within the rift zone depend on the original crustal thickness, t_c, (and β). The initial response to stretching is uplift if t_c is less than some critical value t_c^*, and subsidence if t_c is greater than t_c^*. This critical thickness depends on the assumed values of several variables whose values are imprecisely defined; specifically

$$t_c^* = a\alpha T_1 \rho_o / [2(\rho_o - \rho_c)], \qquad (1)$$

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where α is the coefficient of thermal expansion, ρ_o is the density of mantle material at T = 0°C, and ρ_c is the density of crustal rocks at T = 0°C. For a given, fixed, value of t_c, variations in lithosphere thickness and/or the density difference ($\rho_o - \rho_c$) could allow $t_c < t^*$ in one location and $t_c > t^*$ in another, resulting in initial uplift in the former case but subsidence in the latter. A resonable estimate for t_c is ∿18 km (McKenzie, 1978). However allowing for uncertainties of the parameters in Equation (1), t^* could have any value between 9 km and 72 km. Since normal values of t_c for continental crust also lie within this range graben formation by lithosphereic extension could be associated with either initial uplift or subsidence.

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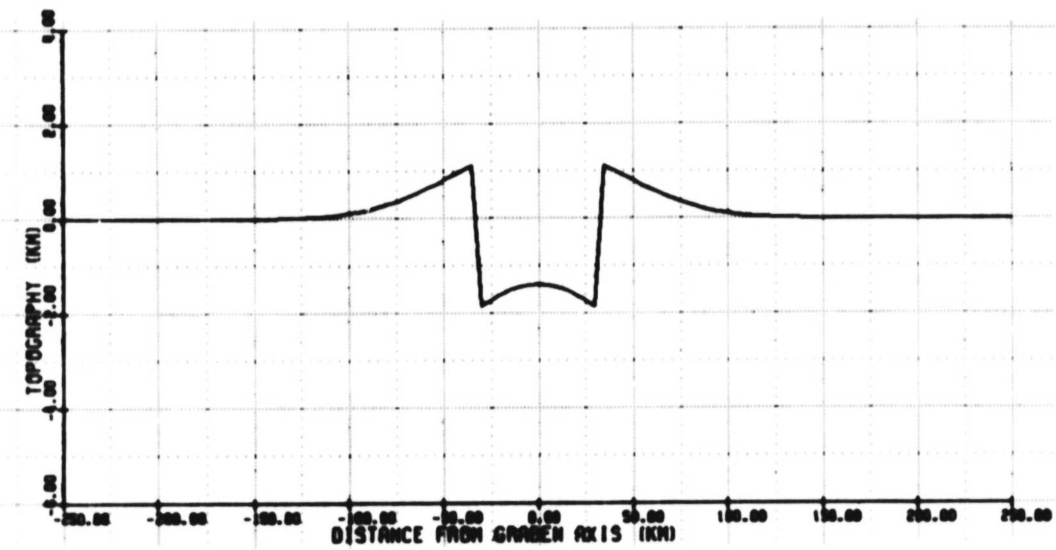

Model Graben Profile at t = 31 Ma due to horizontal extension and thermal diffusion. Model parameters: W = 30 km; β = 2; t_c = 30 km; a = 250 km.

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