Are Transform Faults Thermal Contraction Cracks?

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The oceanic lithosphere cools as it moves away from the ocean ridge where it was formed. Thermal stresses will be associated with this cooling unless the stresses are relieved by plastic flow. Studies of the bending of the lithosphere at Hawaii and at ocean trenches show that the upper 25 km of the lithosphere behaves elastically on geological time scales. It is concluded that thermal stresses will develop at temperatures below 300°C. It is determined that the thermal stresses in the cooling oceanic lithosphere are as high as 4 kbar. These stresses are sufficiently large to fracture the lithosphere. The thermal stresses in the oceanic lithosphere give a bending moment. For a ridge segment with a length of less than about 100 km the thermal stresses can be relieved by bending of the lithosphere; for longer ridge segments the hydrostatic restoring force restricts bending. Transform faults and associated fracture zones are universal features of the ocean ridge system. They exhibit the graben structure associated with crustal extension. It is suggested that transform faults and fracture zones are contraction cracks that relieve the thermal stresses in the cooling lithosphere.

In the context of plate tectonics the outer shell of the earth, the lithosphere, forms rigid plates in relative movement with respect to each other. The lower boundary of the lithosphere is an isotherm T_L [*Turcotte and Oxburgh*, 1972]. At temperatures greater than T_L, mantle rocks behave like a fluid (either Newtonian or non-Newtonian) on geological time scales, and at temperatures below T_L the rocks of the mantle and crust are essentially rigid at low stress levels. A reasonable value for T_L would be 1000° ± 200°C, and the average thickness of the lithosphere would be about 100 km.

Surface evidence indicates that the lithospheric plates preserve their shapes on geological time scales. Evidence is the matching of shape across the Atlantic Ocean and the linearity of such features as the fracture zones on the floor of the Pacific. Lateral motion between plates takes place on major faults like the San Andreas. For strike slip earthquakes, well-defined fractures extend to depths of about 15 km. Apparently, the colder part of the lithosphere behaves like an elastic brittle solid. Studies of the bending of the lithosphere at Hawaii [*Walcott*, 1970] and at trenches [*Hanks*, 1971] show that the oceanic lithosphere exhibits elastic behavior on the time scale of millions of years. By using their values for the flexural rigidity a thickness of 25 km is obtained. Although the upper few kilometers may be sufficiently fractured to be weak in tension, for simplicity it is assumed that the upper 25 km of the oceanic lithosphere behaves like an elastic brittle material on geological time scales. It is also concluded that in the depth range 25–100 km the oceanic lithosphere exhibits plastic behavior on geological time scales. At low stress levels this region is rigid and can act as a stress guide, but large stresses (say, above 10 bar) are relieved by plastic deformation. This definition and this division of the lithosphere (see below) are valid only on the average and are in considerable error in regions of high heat flow.

Depth Range, km	Deformation Mechanism
0–25	elastic brittle
25–100	plastic
100+	fluid

Since the upper 25 km of the lithosphere behaves elastically on geological time scales, this region is subject to thermal stresses. *Turcotte and Oxburgh* [1973] showed that the magnitude of these thermal stresses is sufficient to fracture the lithosphere. In this paper the thermal stresses in the cooling lithosphere adjacent to an ocean ridge are examined in some detail. It is suggested that transform faults and the associated fracture zones are contraction cracks resulting from the cooling of the lithosphere.

Thermal Stresses and Bending Moments

New oceanic lithosphere is created at an ocean ridge and cools as the surface plate convects at a velocity u_0 away from the ridge. A two-dimensional ridge model is illustrated in Figure 1. The ridge axis is at $x = 0$; spreading takes place in the x direction, y is measured along the ridge, and z is measured downward. The temperature distribution for such a model has been considered by many authors. On the basis of the analysis of *Turcotte and Oxburgh* [1967, 1969] it is assumed that the temperature as a function of x and z is given by

$$\frac{T - T_S}{T_L - T_S} = \frac{2}{\pi^{1/2}} \int_0^{z/2[u_0/\kappa(x+x_0)]^{1/2}} \exp(-t^2)\, dt = \operatorname{erf}\left\{\frac{z}{2}\left[\frac{u_0}{\kappa(x + x_0)}\right]^{1/2}\right\} \quad (1)$$

where T_S is the surface temperature and x_0 is chosen to give the correct thermal structure at the ridge crest. For numerical examples the following values are assumed: T_S = 0°C, T_L = 1200°C, u_0 = 1 cm/yr, and $\kappa = 10^{-2}$ cm²/s. The magnetotelluric studies of *Hermance and Grillot* [1970] give T = 980°C at z = 17.5 km beneath Iceland ($x = 0$); the corresponding value for x_0 is 27.5 km.

In order to determine thermal stresses from the temperatures given in (1), boundary conditions are required. Initially, it is assumed that bending of the lithosphere does not occur and that no strain is allowed in the y direction. Since the ridge crest is a zone of weakness, the net force in the lithosphere in the x direction should be small. To simplify the analysis, it is assumed that $\sigma_{xx} = 0$ (deviations from the hydrostatic state are considered); this is not strictly true, since there will be differential thermal stresses in the x direction. The error in the calculated maximum thermal stress due to this assumption is estimated to be about 10%. It is a good approximation to assume $\sigma_{zz} = 0$.

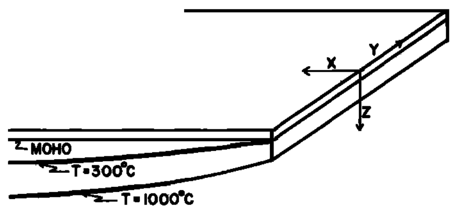

Fig. 1. Model for the cooling lithosphere adjacent to an ocean ridge.

The temperature at the ridge crest is $T_0(z)$. As the lithosphere cools, the temperature is $T(x, z)$. The thermal stress in the y direction on the above assumption is given by

$$\sigma_{yy} = \alpha_l E(T_0 - T) \tag{2}$$

where α_l is the coefficient of linear thermal expansion and E is Young's modulus. This relation is valid as long as the rocks of the lithosphere behave like an elastic material. If the thermal stresses are relieved by plastic flow, then (2) is not valid. The elastic behavior of the lithosphere can be used to determine the maximum temperature for which (2) is valid. Studies of the bending of the oceanic lithosphere at Hawaii [*Walcott*, 1970] and at ocean trenches [*Hanks*, 1971] give a flexural rigidity $D = 2 \times 10^{30}$ dyn cm. By using the definition of the flexural rigidity, $D = Eh^3/12(1 - \nu^2)$, where ν is Poisson's ratio and h is the thickness of the elastic plate; for $E = 1.5 \times 10^{12}$ dyn/cm^2 and $\nu = 0.25$, the thickness h is 25 km. If it is assumed that the surface heat flow in these regions is near 1 μcal/cm^2 s and $k = 8 \times 10^{-3}$ cal/cm s °C, the temperature T_p at a depth of 25 km is 300°C.

We conclude that at temperatures below $T_p = 300$°C the lithosphere behaves elastically on time scales of millions of years but at higher temperatures plastic yielding takes place, thus relieving large elastic stresses. Thermal stresses can develop on geological time scales only at temperatures below $T_p = 300$°C, and (2) is valid in the range $0 < T_0 < T_p$. As the cooling lithosphere cools through the 300°C isotherm, thermal stresses will develop, and therefore

$$\sigma_{yy} = \alpha_l E(T_p - T) \tag{3}$$

if $T < T_p < T_0$. It is expected that the critical temperature T_p will be time dependent; for time scales of 10^6–10^9 yr the temperature may be somewhat less than 300°C. At the present time there are insufficient data to evaluate this time dependence. Since the time scale for the thickening of the thermal boundary layer near a ridge is similar to that considered by *Walcott* [1970] and *Hanks* [1971], the value $T_p =$ 300°C should be applicable.

By using the temperature profile given in (1) it is found that $T_0 = 300$°C at $z = 4.12$ km. At the ridge crest the lithosphere behaves elastically to this depth. This finding is also in good agreement with the maximum depth of ridge seismicity. In order to determine thermal stresses we require the coefficient of linear thermal expansion α_l and Young's modulus E for the lithosphere. We assume that the lithosphere is made up of a basaltic crust with a thickness of 4.12 km and a pyrolite mantle [*Ringwood*, 1966]. For the basaltic crust we take $\alpha_l = 0.7 \times 10^{-5}$ °C^{-1} and $E = 10^{12}$ dyn/cm^2 [*Clark*, 1966]. For the pyrolite mantle we take $\alpha_l = 10^{-5}$ °C^{-1} and $E = 1.7 \times 10^{12}$ dyn/cm^2. Since the product $\alpha_l E$ differs in the crust and the mantle, the lithosphere behaves like a bimetallic strip. Differential thermal stresses develop at the Moho.

The calculated thermal stresses are given in Figure 2 as a function of depth z for several values of x. Because of the assumption of no strain in the y direction, all thermal stresses are tensions. The thermal stress at the surface ($z = 0$) is zero, since the temperature remains constant. The thermal stress increases with depth in the crust. Because of the larger $\alpha_l E$ product in the mantle the thermal stresses in the mantle are larger than those in the crust. The freezing-in of thermal stress in the mantle as the thermal boundary layer develops is clearly illustrated, and the thermal stress decreases with depth. Maximum thermal stresses of about 4 kbar develop. The maximum thermal stress as a function of x is given in Figure 3. The maximum thermal stress occurs at the depth where $T_0 = T_p$.

Also shown in Figure 2 is the hydrostatic pressure as a function of z. If it is postulated that a fracture will initiate when the tension stress exceeds the hydrostatic pressure, the fracture will be expected to initiate at the Moho 20 km from the ridge crest. From this point the fracture will propagate upward into the crust and downward into the mantle. Once it is initiated, it is expected that thermal stresses will continue to be relieved along the fracture zone.

The thermal stresses will generate a thermal bending moment on the lithosphere. The thermal bending moment per unit width is given by

$$M_{th} = \int_0^{z_t} (\sigma_{yy} - \bar{\sigma}_{yy}) z \, dz \tag{4}$$

where $\bar{\sigma}_{yy} = (1/z_t)\int_0^{z_t} \sigma_{yy} \, dz$, where z_t is the thickness of the elastic part of the lithosphere, in which there are thermal stresses and $\bar{\sigma}_{yy}$ is the mean value of the thermal stress in the lithosphere. If M_{th} is positive, the lithosphere will be depressed, and if M_{th} is negative, the lithosphere will be arched. The thermal bending moment is plotted against z in Figure 3. From $x = 0$ to about 200 km the thermal bending moment is negative. The maximum tensile thermal stress is near the base of the elastic layer, and the tendency is to bend the lithosphere convex up. As the thermal boundary layer develops further, the maximum tensile thermal stress is near the top of the elastic layer, and the tendency is to bend the lithosphere concave up. It should be emphasized that no bending of the lithosphere has been allowed in determining the above thermal stresses and thermal bending moments.

If the thermal stresses are relieved by fracture, the net force

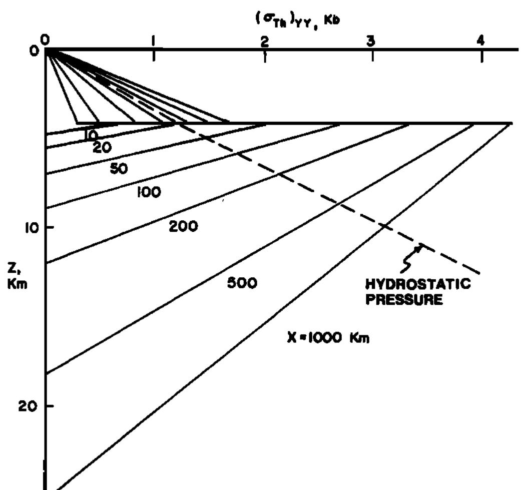

Fig. 2. Thermal stresses in the cooling lithosphere as a function of depth at several distances from the ridge crest.

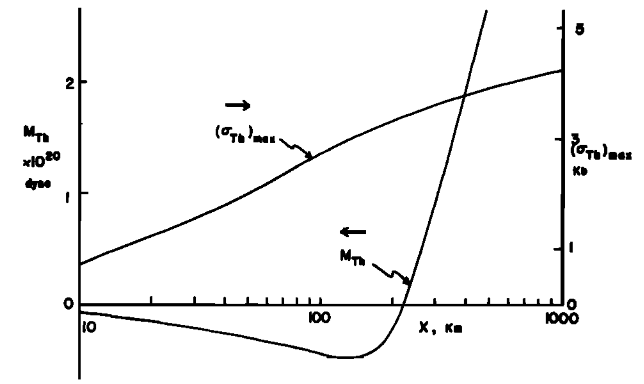

Fig. 3. Thermal bending moment and maximum thermal stress in the cooling lithosphere as a function of the distance from the ridge crest.

in the y direction will be zero: $\bar{\sigma}_{yy} = 0$. This will drive the upper part of the lithosphere into compression and is one of the explanations given by *Sykes and Sbar* [1973] for the observed predominance of compressive stresses in the oceanic lithosphere.

Bending of the Lithosphere

Thermal bending stresses in the lithosphere may cause the lithosphere to bend. To consider this effect, it is necessary to allow a vertical displacement of the lithosphere w. It is assumed that the lithosphere is broken into a series of segments of length l. We will argue in the next section that the termination of each segment is a transform fault. To simplify the analysis, only thermal stresses in the y direction are considered, and variations of the thermal bending moment in the x direction are neglected; it is assumed that $w = w(y)$.

The bending of the lithosphere is restricted by a hydrostatic head [*Walcott,* 1970]. The vertical displacement of the lithosphere satisfies

$$D\, d^4w/dy^4 + (\rho - \rho_w)gw = 0 \qquad (5)$$

where ρ_w is the density of water, ρ is the density of the lithosphere, g is the acceleration of gravity, and D is the flexural rigidity previously defined. Introducing the flexural parameter $f = [4D/(\rho - \rho_w)g]^{1/4}$ into (5) gives

$$d^4w/dy^4 + (4/f^4)w = 0 \qquad (6)$$

A number of boundary conditions could be applied at the edge of the ridge segment. A solution will be obtained for a free edge; that is, the applied bending moment and shear stress at the edges of the plate are zero, so that $D\, d^2w/dy^2 = M_{th}$ and $d^3w/dy^3 = 0$ at $y = \pm l/2$. The thermal bending moment enters the problem through the boundary conditions. Without the hydrostatic restoring force the plate would have a uniform curvature M_{th}/D. With the hydrostatic restoring force the solution of (6) for the vertical displacement is given by

$$w = \frac{M_{th}f^2}{2D}\left[\left(\frac{\cos l/2f \sinh l/2f + \sin l/2f \cosh l/2f}{\sin l/2f \cos l/2f + \sinh l/2f \cosh l/2f}\right)\cdot \sin\frac{y}{f}\sinh\frac{y}{f} + \left(\frac{\cos l/2f \sinh l/2f - \sin l/2f \cosh l/2f}{\sin l/2f \cos l/2f + \sinh l/2f \cosh l/2f}\right)\cdot \cos\frac{y}{f}\cosh\frac{y}{f}\right] \qquad (7)$$

and the bending moment in the plate is given by

$$M = M_{th}\left[\left(\frac{\cos l/2f \sinh l/2f - \sin l/2f \cosh l/2f}{\sin l/2f \cos l/2f + \sinh l/2f \cosh l/2f}\right)\cdot \sin\frac{y}{f}\sinh\frac{y}{f} - \left(\frac{\cos l/2f \sinh l/2f + \sin l/2f \cosh l/2f}{\sin l/2f \cos l/2f + \sinh l/2f \cosh l/2f}\right)\cdot \cos\frac{y}{f}\cosh\frac{y}{f} + 1\right] \qquad (8)$$

In the limit $l/f \to 0$, $w \to M_{th}y^2/2D$, and $M \to 0$; i.e., the plate bends freely and relieves the thermal bending moment.

As a particular example we consider the bending of the lithosphere 100 km from the ridge crest. The thermal bending moment is $M_{th} = -4.5 \times 10^{19}$ dyn. The thickness of the elastic lithosphere is $h = 8.9$ km, $D = 8.5 \times 10^{28}$ dyn cm, and $f = 36$ km. The vertical displacement of the plate w from (7) is given as a function of y in Figure 4 for $l = 72$, 144, and 216 km. The maximum vertical displacement is of the order of 40 m. The bending moment at the center of the ridge segment $y = 0$ from (8) is plotted against the length of the ridge segment l in Figure 5. For a length $l = 72$ km the hydrostatic restoring force is unimportant, and the lithosphere is free to bend; the bending relieves the thermal stresses, so that the bending moment in the lithosphere is small. As the length of the ridge seg-

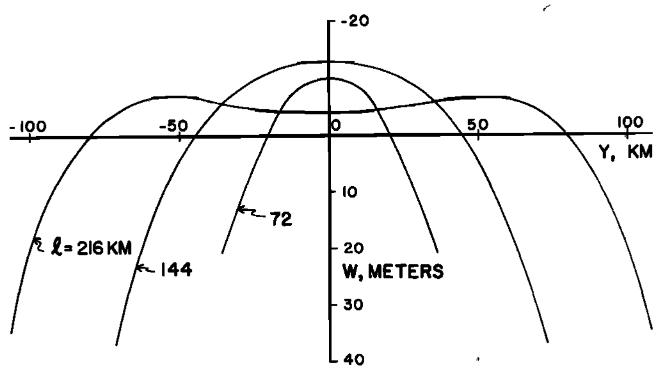

Fig. 4. Vertical displacement of the lithosphere due to a thermal bending moment $M_{th} = -4.5 \times 10^{19}$ dyn for several lengths of the ridge segment.

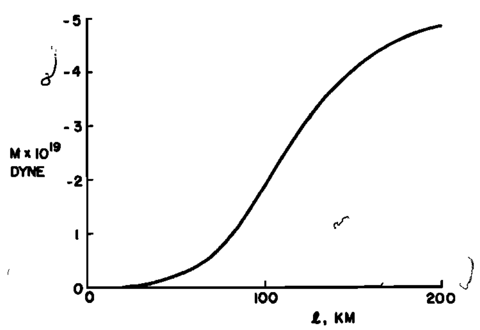

Fig. 5. Bending moment at the center of the ridge segment as a function of the length of the ridge segment.

ment is increased to l = 144 and 216 km, the hydrostatic restoring force inhibits bending, and the bending moment in the elastic lithosphere increases.

Once a thermal fracture occurs, some of the thermal stress is relieved. Without bending, the near-surface region goes into compression with tension at depth. Ideally, the integrated thermal stress will be zero. However, the tensional stresses near the Moho may still be sufficient to cause thermal fractures. These tensional stresses can only be relieved by the bending of the lithosphere. Since bending can occur only for ridge segments shorter than about 100 km, the spacing of the tension fractures should be between 100 and 200 km. This interval is in good agreèment with the spacing between transform faults, as will be discussed in the next section.

TRANSFORM FAULTS

The role of transform faults in offsetting segments of the ocean ridge system was first suggested by *Wilson* [1965]. It is generally accepted that the major fracture zones of the ocean basins are extensions of these transform faults. Although topography and seismicity can be used to locate transform faults, offsets in magnetic anomalies seem to be the best evidence for the presence of transform faults and associated fracture zones.

By using the worldwide studies of magnetic anomalies the length of ridge segments as a function of offset is given in Figure 6. There are clearly two types of ridge segments: type 1, where the offsets at the end of the ridge segment tend in the same direction (Figure 7, top), and type 2, where the offsets tend in opposite directions (Figure 7, bottom). There seems to be little correlation between the length of the segment and the offset. Transform faults seem to be a universal feature of the ocean ridge system. In Figure 8 the distribution of ridge segment lengths is given. Most ridge segments have a length between 50 and 200 km; the mean length of a ridge segment is 250 km. There is no clear distinction between type 1 and type 2 transforms. The length of the segments appears to be independent of whether spreading is nearly normal to the ridge trend (thus favoring type 2 ridge segments) or is oblique to the ridge trend (thus favoring type 1 ridge segments). Since transforms with small offsets are difficult to discover, the mean length will certainly decrease as more accurate studies of magnetic anomalies and other means of identifying transform faults become available. It should be noted that transform faults are also observed in convection in lava pools [*Duffield,* 1972] and in laboratory experiments using wax [*Oldenburg and Brune,* 1972].

Several suggestions for the cause of transform faults have been made. A number of authors [*Menard and Atwater,* 1968,

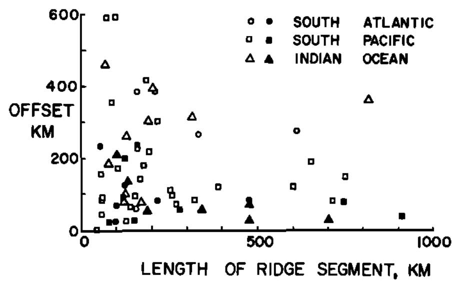

Fig. 6. Sum of the offsets at the ends of a ridge segment as a function of the length of the ridge segment; the open symbols represent offsets in the same direction (type 1), and the solid symbols represent offsets in the opposite direction (type 2).

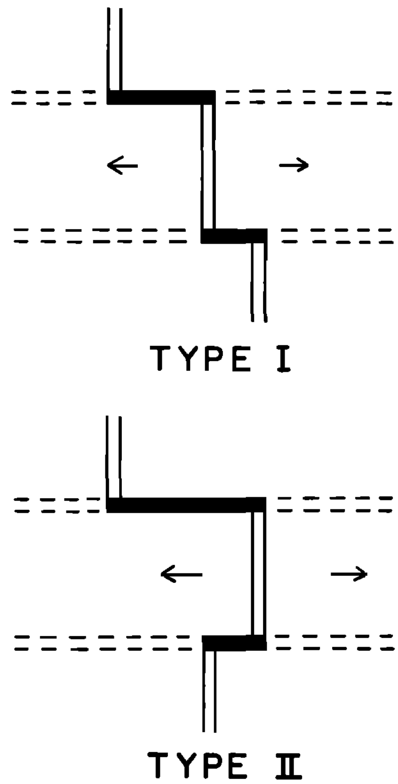

Fig. 7. Two types of ridge-transform segments; the double line represents the ridge segment, the solid line represents the transform fault, and the double dashed line represents the fracture zone.

1969; *Van Andel et al.,* 1969, 1971] associate transform faults with small changes in spreading direction. This suggestion does not seem to be consistent with the observation that transform faults are a universal feature of ocean ridge systems. *Lachenbruch and Thompson* [1972] have suggested that the ridge-transform system reduces the resistance to sea floor spreading and it, rather than oblique ridges, is therefore observed. There is no reason to believe that such a minimum-resistance principle is applicable to a nonlinear process like sea floor spreading. However, more importantly, this explanation is not applicable to ridges where spreading is normal to the ridge axis; observations indicate that transform faults are present in these regions also.

Since transform faults are associated with the formation of a solid elastic layer due to the cooling of a nonelastic 'fluid,' it is expected that thermal contraction cracks will be present. This explains the distribution of transform faults throughout the worldwide ridge system and also their presence in lava convection and in laboratory studies using wax. The analysis given in this paper shows that the expected thermal stresses are sufficient to cause contraction fracturing of the lithosphere. The analysis of the bending of the lithosphere shows that bending can relieve thermal stresses for ridge segments shorter than about 100 km.

Thermal contraction also explains the structural features of transform faults. Transform faults have a rift valley character; the sides appear to have rifted apart [*Collette and Rutten,* 1972]. As the ridge segment thermally contracts through cooling, the gap appears at the transform fault. For a 150-km ridge segment and a ΔT of 200°C an extension of about 300 m would be expected across the transform.

Clearly, the analysis given in this paper does not explain all the aspects of the ridge-transform system. A complete

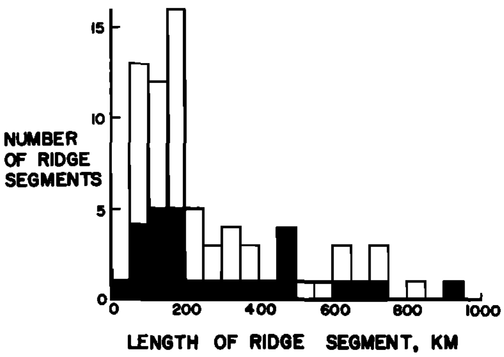

Fig. 8. Distribution of ridge segment lengths in 25-km increments; the open lines represent type 1 ridge segments, and the solid lines represent type 2 ridge segments.

analysis would require a three-dimensional study of the stress in the lithosphere including the mechanics of both the transform faults and the ridge crest. Thermal stresses normal to the ridge should be included; it is possible that these stresses are responsible for the 'jumping' of spreading centers. Also, no attempt has been made to explain the presence of ridge offsets. Once the ridge is broken by a thermal contraction crack, small random variations in the spreading rates on the two sides of the ridge will lead to offsets. The difference in spreading rates could be attributed to the details of mass addition at the ridge crest. The initiation of transform faults may be related to the geometry of the initial continental breakup. For example, the transform faults in the Red Sea and Gulf of Aden seem to line up with offsets in the continental coastline.

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