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A REVIEW OF THE STRUCTURE OF MID-OCEAN RIDGE SYSTEMS WITH APPLICATIONS TO THE TECTONIC HISTORY OF THE NORTHEAST PACIFIC

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ABSTRACT

The physical properties of mid-ocean ridge systems are determined primarily through the use of thermal models. It was found that a leaky transform fault could be described by using a similar method. The stability of these ridge systems depends on the stress field that is acting upon them. Mid-ocean ridges can undergo substantial deformation which can be brought about by changes in the stress field. Such an effect is expected to be greater as a ridge gets in close proximity of a trench. In the case of the northeast Pacific, the model of Atwater (1970) for the tectonic history of this region is not compatable with the observed magnetic data, under the assumption that major changes in the stress field should be mirrored in the magnetic anomalies. A newer model in which the East Pacific Rise reaches the trench at 24 myBP is presented here. Such a model requires that the trench remain for another 4 m.y. before converting into a transform fault.

INTRODUCTION

Changes in the direction of sea-floor spreading have been well documented (eg. Menard and Atwater, 1968; Francheteau et al., 1970). Such changes have been attributed to a shift in the pole of rotation of the plates and have been identified on the basis of bends in the trends of fracture zones. This is based upon the assumption that these fracture zones were formed parallel to the direction of spreading and therefore represent small circles about the pole of rotation of the plates (Morgan, 1968). Furthermore, the recognition of a systematic series of magnetic lineations corresponding to reversals in the magnetic field of the earth has resulted in the ability to deduce the spreading history of the ocean floor (Vine and Matthews, 1963). Consequently, it has been found that changes in the rate of sea-floor spreading occurred existed in the past (Vogt et al., 1969). In most cases, these changes have been correlated to various orogenic events and are believed to be somewhat related to them. However, while the general kinematics of these deviations in constant motion are relatively known (eg. Atwater, 1970; McKenzie and Morgan, 1969), the physical properties are still relatively questionable.

In general, papers concerned with the physical mechanisms of mid-ocean ridges have dealt mostly with the steady-state case in which physical equilibrium is assumed (eg. McKenzie, 1967; Sleep, 1975). And while newer models which investigate nonequilibrium situations are currently being developed, it may prove to be useful to analyze the dominant factors which affect the equilibrium by looking at the spreading history in an area which has undergone significant tectonic readjustment. By doing so, it will hopefully provide some constraints upon the physical properties of mid-ocean ridge systems.

Therefore, this paper will essentially be divided into two general sections. We will begin with a brief review of some of the current ideas on the structure of mid-ocean ridge systems (ridges and transform faults) and will end by looking at the spreading history of a particular area. In this case, the area chosen is the northeast Pacific basin (Fig. 1). This region is of particular interest because it has been hypothesized

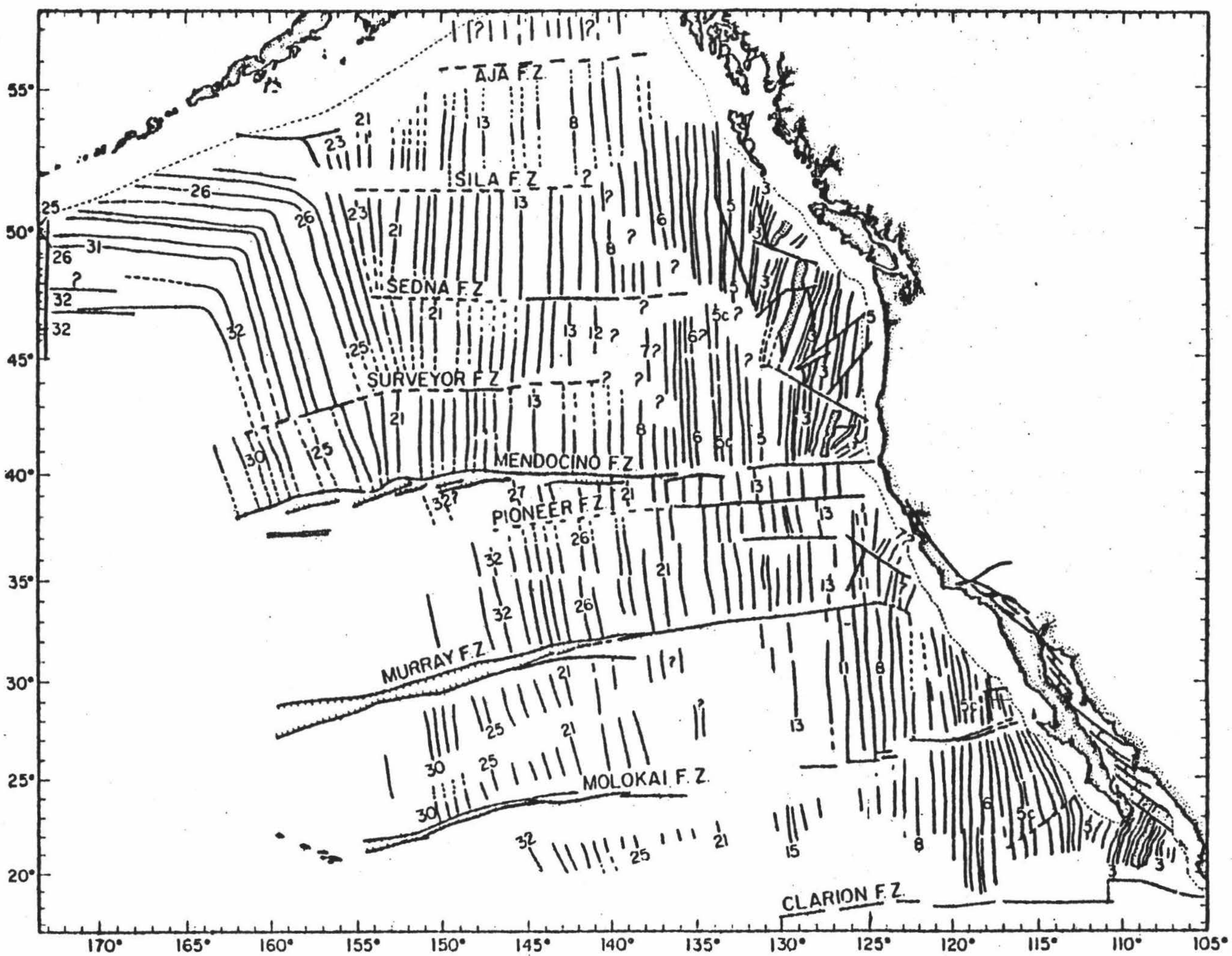


Fig. 1. Magnetic anomalies in the northeast Pacific. Anomalies as interpretted by Atwater and Menard (1970) with the numbering sequence following that of Heirtzler et al. (1968) Figure from Atwater and Menard (1970).

that three different mid-ocean ridges have been subducted since the Cretaceous period. These ridges being the Kula-Pacific ridge, the Kula-Farallon ridge, and the East Pacific Rise (EPR) (Atwater, 1970). This paper will focus on the last one mentioned, the East Pacific Rise.

A major problem in doing such a study in which a ridge gets subducted is that half the spreading history is destroyed when the leading plate (in this case either the Kula or Farallon) goes down the trench. Without the constraints that these plates would have provided, the tectonic history and therefore also the physical mechanisms involved may not be fully determined.

THERMAL STRUCTURE OF MID-OCEAN RIDGES

The physical mechanisms of a mid-ocean ridge (MOR) have been the subject of a great deal of speculation. The primary purpose of presenting the following thermal models for the formation of oceanic lithosphere is to determine the shape and location of the magma chamber and to also get some insight into the dominant factors involved. The traditional method in analysing the process that occur under a MOR is through the use of a theoretical thermal model in which the oceanic lithosphere is approximated by a thin slab of some thickness λ, overlying a hot asthenosphere (McKenzie, 1967). From this idealized case, a thermal structure of the lithosphere as it moves away from the ridge can be obtained. McKenzie (1967) pointed out that the heat flow equation in this case was

$$\rho C_p \left[\frac{\partial T}{\partial t} + \vec{v}\cdot\vec{\nabla} T \right] = \kappa \nabla^2 T + H \qquad (1)$$

where, ρ = density
$\vec{v}$ = velocity
T = temperature
κ = thermal conductivity
H = internal heat production in the slab

If one only considers the steady-state case with no internal heat production (ie. no radioactive heat source), then the generalized equation reduces to

$$\rho C_p \, \vec{v}\cdot\vec{\nabla} T = \kappa \nabla^2 T \qquad (2)$$

Since spreading is normal to the strike of the ridge, then by specifying the coordinate system by $\hat{k}$ being positive downwards and spreading in the $\hat{i}$ direction, equation 2 reduces to

$$\rho C_p \, v \frac{\partial T}{\partial x} = \kappa \left[\frac{\partial^2 T}{\partial x^2} + \frac{\partial^2 T}{\partial z^2} \right] \qquad (3)$$

The solution given by McKenzie for the boundary conditions, $T(0)=0$ and $T(\lambda)=T_\lambda$ is

$$T(z,x) = \frac{T_\lambda z}{\lambda} + \sum^{\infty} a_n \exp\left[\left(R - \sqrt{R^2 + n^2\pi^2}\right) \frac{x}{\lambda} \right] \sin\left(\frac{n\pi z}{\lambda}\right)$$

where R = Rayleigh number
$= \rho C_p v\lambda / 2\kappa$

and a_n is determined by the other boundary conditions.

However, the previous idealized model is lacking in a couple of aspects. Primarily, the deficencies are that a mathematical singularity exists directly over the mid-ocean ridge (Lubimova and Nikitina, 1975) and secondly, that the thickness of the slab is not independently derived. Parker and Oldenburg (1973) modified McKenzie's equations to include a term which took into account the latent heat of fusion (heat due to the liquid-solid phase transition) and letting the thickness of the lithosphere be determined by the depth to the solidus temperature isotherm. The new heat flow equation and boundary conditions are thus

$$\vec{n}\cdot\vec{v}\rho L = -\kappa\left|\vec{\nabla}T\right| \tag{5}$$

where L = latent heat of fusion
$\vec{n}$ = outward normal to the interface (lithosphere-asthenosphere)

$$-\kappa\frac{\partial T}{\partial x} = \rho v\ [\ L + C_p\ (\ T_\lambda - T_f(z))] \tag{6}$$

and T_λ = temperature at the base of the lithosphere
T_f = final temperature to which the magma cools before the next episode of intrusion

Another refinement was made by Sleep (1974, 1975) in which he changed the boundary conditions to take into account differences in the horizontal heat flux with depth. The new heat flow equation is now modified to include a more precise definition of this effect.

$$\begin{aligned} S &= \text{horizontal heat flux} \\ &= -\kappa\frac{\partial T}{\partial x} + vC_v\ (T - T_\lambda z/\lambda) \\ &= \gamma T_\lambda\ (\lambda - z)\ vC_v/\lambda \end{aligned} \tag{7}$$

where γ = coefficient equal to 1 minus the ratio of adiabatic to thermal gradient

A solution to the new heat flow equation can be found in Sleep (1974).

The differences in the heat conduction with depth is due to the material being below or in the zone of melt-crystalline mush segregation (Sleep, 1975).

Finally, a numerical model devised by Kusznir and Bott (1976) has been used which determines the temperature distribution over a finite difference grid. The advantages of this model ~~is~~ are that if a fine enough grid system is utilized, then the shape of the hypothesized magma chamber can be better constrained. Furthermore, a refinement of the distribution of the latent heat effect was also carried out.

So far, just the methods have been presented. As can be seen, several refinements have been made over the past ten years with varying degrees of difficulty and refinement. To fully understand the effect of each improvement, a comparison of the results is needed. In each case, the following isotherms were calculated under different spreading conditions and physical parameters. The following isotherms taken from the various papers are shown in figure 2.

In interpretting these diagrams, one must keep in mind the location of the isotherm which marks the boundary of the solidus. This is the boundary between the lithosphere and the asthenosphere. As can be seen, there is no complete agreement on what ~~the~~ exact temperature this should be.

In the case of fast versus slow spreading, it can be seen that any isotherm for the faster spreding ridge is closer to the top of the lithosphere for a particular distance away from the ridge. This reflects upon the net effect of horizontal heat conduction as opposed to vertical heat conduction. For the slower spreading case, the former has a more significant effect. This has led to the conclusion that for slow enough spreading (<0.9 cm/yr, Sleep, 1975; < 0.5 cm/yr, Kusznir and Bott, 1976) lateral heat conduction is sufficient ~~enough~~ to freeze the molten material at the base of the lithosphere under the mid-ocean ridge. This may explain why very slow spreading ridges are not found in the oceans.

Based upon the theoretical thermal structure, it has been proposed that the oceanic lithosphere should thicken with age. This thickening excluding the addition of sediments can be seen in the position of the wet or dry solidus isotherm. As the lithosphere gets older, the position of the solidus increases in depth. This conclusion is also supported by seismic surface wave dispersion in paths passing across the Pacific Ocean basin (Leeds et al., 1974). Unfortunately, several discrepencies exist. Primarily, seismic evidence for such a thickening effect indicates that the lithosphere does not thicken as fast as expected. Crough (1977) investigated this problem by doing a more detailed analysis of the thermal

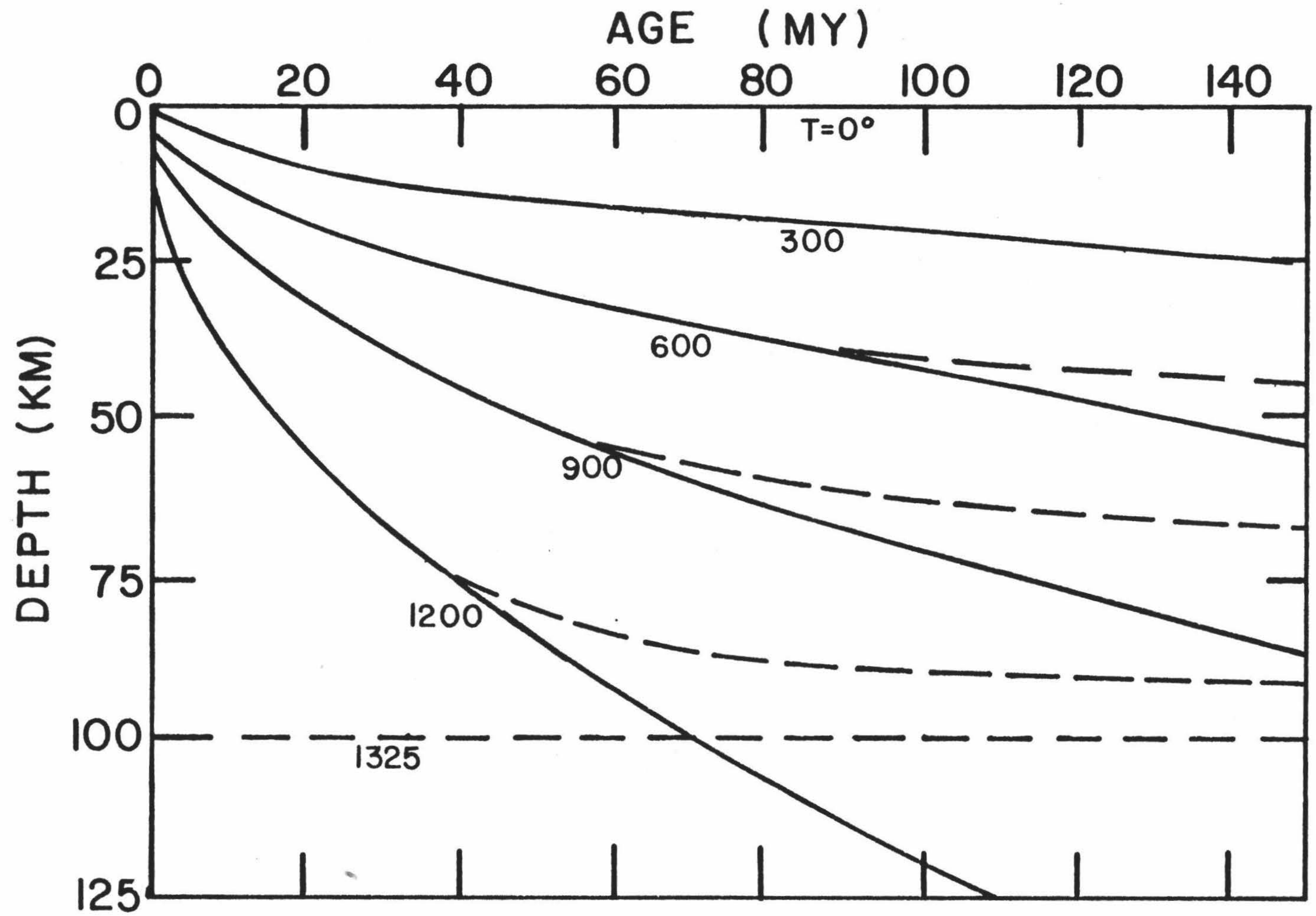


Fig. 2a. Isotherms showing the difference between the flat slab model (dotted lines) and that of Parker and Oldenburg (solid lines). The slab model is for a thickness of 100 km. Note the similarites up to around 50 m.y. in age. Figure from Forsyth (1977).

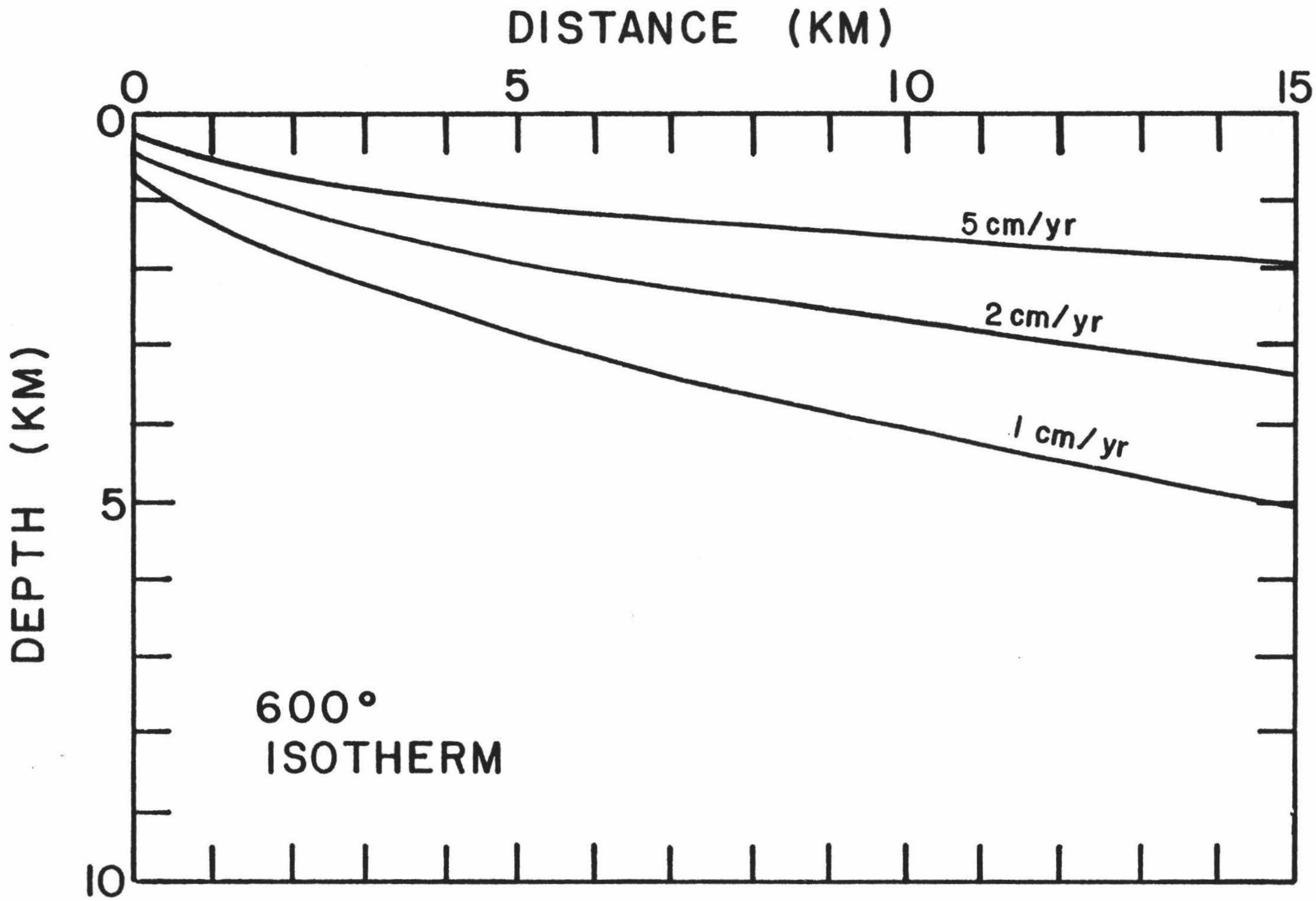


Fig. 2b. Location of the 600°c isotherm for different spreading rates. Note that the location of the isotherm is closer to the surface for faster spreading rates. Figure adapted from that of Sleep (1975).

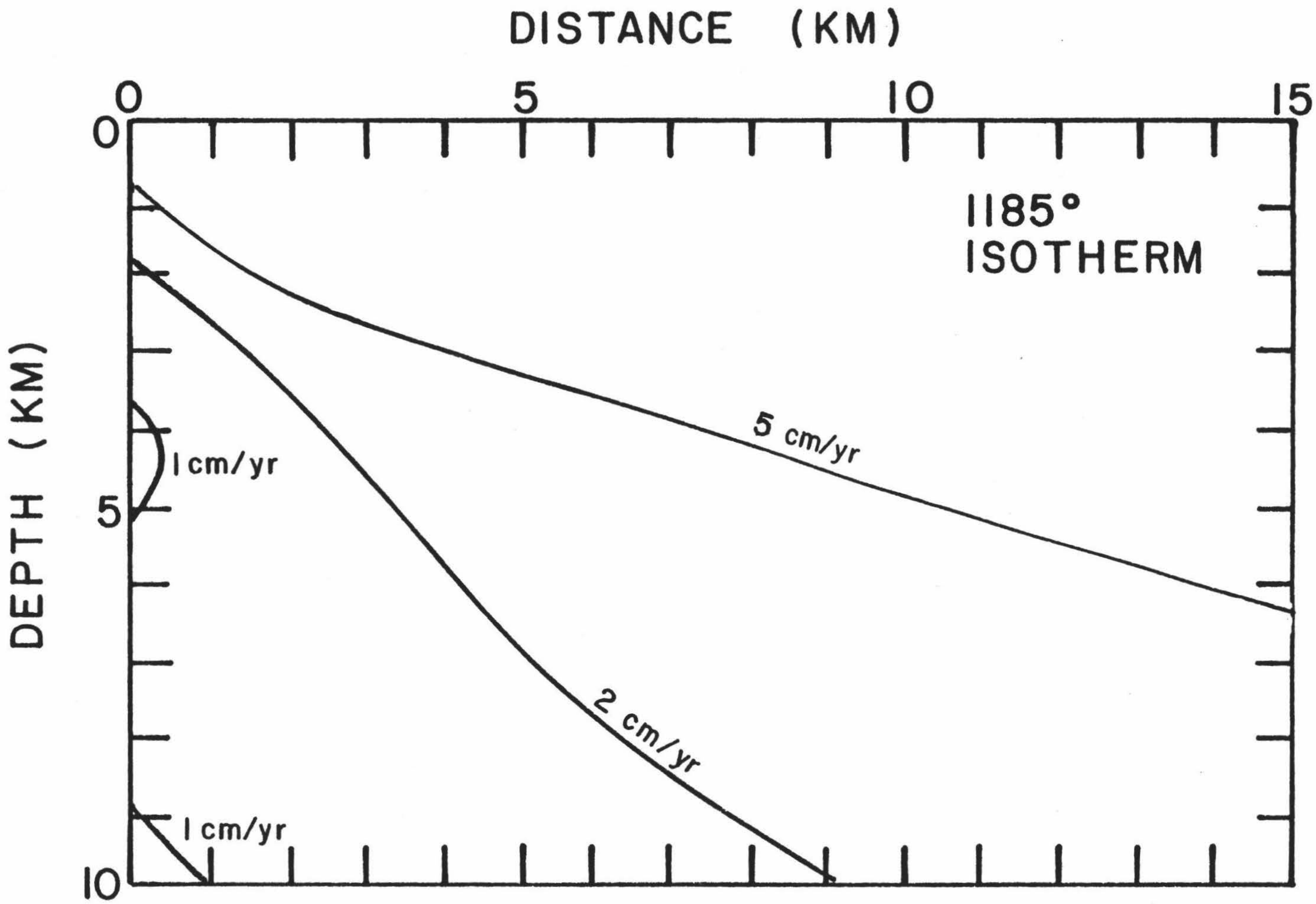


Fig. 2c. 1185°c isotherm taken from Sleep (1975). This essentially shows the location of the solidus. Note the location for spreading rate of 1 cm/yr.

structure of the lithosphere. This study included the contribution of several factors omitted in earlier studies. These ~~being~~ are the internal heat effect (H in the McKenzie, 1967 heat flow equation), partial melt in the asthenosphere, temperature gradient in the asthenosphere and a non-zero lithospheric thickness at the ridge crest. It was found that while the effect of each of these factors is small, they all contribute to the delaying of the cooling of the oceanic lithosphere.

PETROLOGY OF MID-OCEAN RIDGES

The petrological nature of a mid-ocean ridge is not fully understood, although several hypotheses do exist. It is known that the dominant rock type from the top of the oceanic lithosphere (excluding sediments) is an olivine tholeiitic basalt. It is characterized by relatively low K and $^{87}Sr/^{86}Sr$ and by relatively high concentration of Al_2O_3. As can be seen from table 1 (Melson et al., 1976), there is very little deviation in the major element chemistry of rocks associated with different spreading rates with the exception of slightly higher FeO (total Fe) and TiO_2 for rocks from the Pacific as compared to that from the Atlantic. Based upon this, several authors have concluded that this reflects upon the difference in spreading rates between the Mid-Atlantic Ridge (around 1.5 cm/yr) and the East Pacific Rise (EPR) (around 8.0 cm/yr) (Melson et al., 1976). Similarily, it has been proposed that the depth of magma generation is deeper under the Mid-Atlantic Ridge (MAR) (23 km.) compared to the EPR (16 km.) (Scheidegger, 1973). This was

TABLE 1 Averages for Spreading Centers. N Equals Number of Groups Used for Making Averages. Samples from Fracture Zones, Marginal Basins, Seamounts, and Off-Ridge Intrusions Have Been Deleted

	N	SiO_2	Al_2O_3	FeO*	MgO	CaO	Na_2O	K_2O	TiO_2	P_2O_5
Mid-Atlantic ridge	51	50.68	15.60	9.85	7.69	11.44	2.66	0.17	1.49	0.12
East Pacific rise[1]	38	50.19	14.86	11.33	7.10	11.44	2.66	0.16	1.77	0.14
East Pacific rise[2]	20	50.31	14.95	10.71	7.36	11.73	2.60	0.11	1.61	0.12
Juan de Fuca ridge	18	50.06	14.76	12.03	6.81	11.12	2.72	0.21	1.95	0.16
Indian Ocean centers	12	50.93	15.15	10.32	7.69	11.84	2.32	0.14	1.19	0.10
All spreading centers	101	50.53	15.27	10.46	7.47	11.49	2.62	0.16	1.56	0.13
MORB [*Cann*, 1971]	—	49.61	16.01	11.49	7.84	11.32	2.76	0.22	1.43	0.14
MORB [*Engel et al.*, 1965]	10	49.94	17.25	8.71	7.28	11.68	2.76	0.16	1.51	0.16

[1]Includes Galapagos rise and Juan de Fuca ridge.
[2]Excludes Juan de Fuca ridge.

From Melson et al. (1976)

based upon the use~~d~~ of the plagioclase geothermometer through which the temperature of formation of the equilibrium melt can be deduced. However, all of this has recently been questioned by Bryan et al. (1976) in their analysis of basalts drilled by the Deep-sea Drilling project (DSDP). They found that while the results presented by Melson et al. are in agreement with their results, the deviations in the values of the major elements were not statistically significant to make any concluding statements relating spreading rate with rock composition. Therefore. while petrological data gives some insight into the internal structure of a mid-ocean ridge, it is far from being conclusive.

Another method to help determine the petrological nature of the oceanic lithosphere would be to examine ophiolite sequences as it has been hypothesized that they represent obducted sections of oceanic lithosphere. However, this point will not be pursued any further at this point as there is still considerable debate concerning their origin.

PROCESSES UNDER A MID-OCEAN RIDGE

Based upon the petrological and thermal evidence just presented, the following model for the processes under a mid-ocean ridge has been proposed. Mantle material rising under the MOR reaches a depth of about 175 km. At this point the adiabat crosses the wet solidus ($<0.4\%$ H_2O) and partial melt begins to form (Fig. 3) (Forsyth, 1977). This rise of mantle material may be due to the motion in convection cells under the ridge (Oxburgh and Turcotte, 1968) or may be simply due to the buoyancy of anomalously hotter material. Extensive melting does not occur at this point since the adiabat is still below the dry solidus (Wyllie, 1971). The composition of this source material is the subject of much current debate. One of the leading models calls for a peridotite orgin as oppose to eclogite (Green and Liebermann, 1976; Forsyth, 1977). The other model favors a two layer case with peridotite forming the upper layer and olivine eclogite the lower layer (in the mantle) (Chung, 1976). Eventually, the adiabat crosses the dry solidus and extensive melt forms. In the model presented by Forsyth (Fig. 3), this corresponds to a depth of around 30 km. The region beween the dry and wet solidus forms the low-velocity zone and contains around 3% partial melt (Anderson and Sammis, 1970). Support for this model comes from seismic evidence in terms of shear wave attenuation studies by Solomon (1973). He found

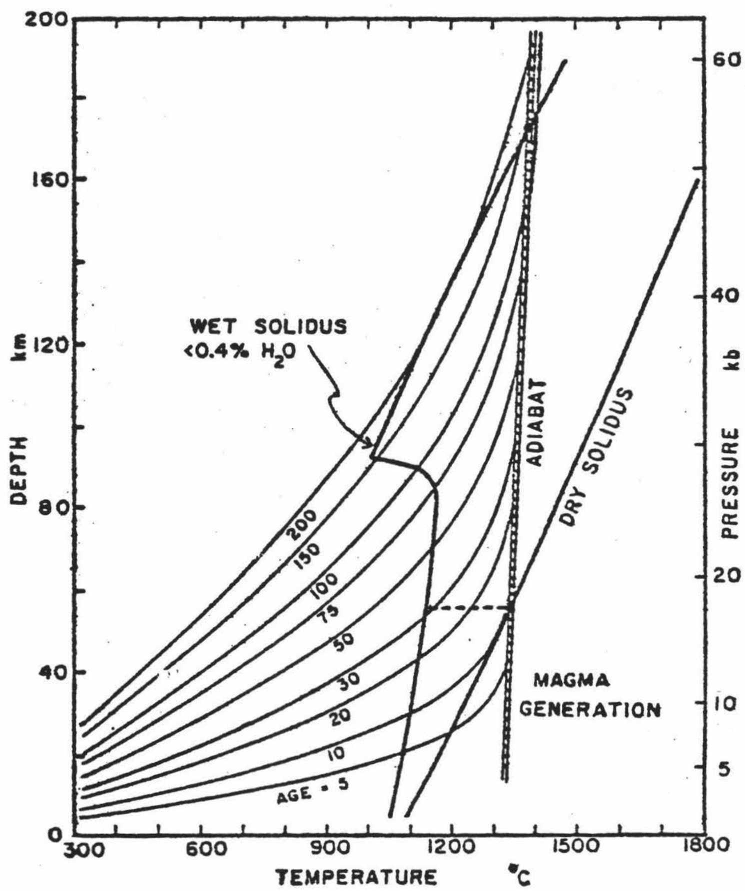


Fig. 3. Figure from Forsyth (1977) which is an adaptation of that of Green and Liebermann (1976). Adiabat is at 1350°c. Wet solidus is intersected at 175 km. depth and the dry solidus at around 60 km. depth.

that a low Q (high attenuation) zone existed under the Mid-Atlantic Ridge between the depth of 50-150 km.

The formation of sufficient melt would then enable the process of filter pressing to occur, thereby separating the crystalline mush from the actual melt (Sleep, 1974). Once enough melt has accumulated, it can then begin to rise by elastic crack propagation or simply by its natural buoyancy. This material eventually reaches a shallow depth under the mid-ocean ridge and can then collect in a magma chamber. As mentioned previously, thermal models which have taken into account the effect of the latent heat of fusion have given some insight into the predicted shape and nature of such magma chamber (Sleep, 1975; Kusznir and Bott, 1976). In Fig. 4 it can be seen that the shape of the isotherms are modified by the different rates of spreading. The shape is also highly dependent of the amount of crystal settling and accretion that takes place. Kusznir and Bott (1976) analyzed the effect of different amount of crystal settling by utilizing a term they called the crystal settling factor, F. This was defined as the ratio of the amount of material accreting to the roof of the magma chamber to the amount falling to the bottom (1:F). The results are shown in Fig. 5.

The very nature of the magma chamber itself is still questionable. Sleep (1975) argues that there would be a chamber consisting of crystal cumulates which then segregate out from the melt (Fig. 6). He therefore questions the possibility of having a completely molten chamber. This argument is also presented by Kusznir and Bott (1976) and by Solomon and Julian (1974). The latter of the two used seismic evidence.

This concludes the section on the structure of mid-ocean ridges. We must now look at the features that also make up a ridge system, this being the transform faults.

PHYSICAL PROPERTIES OF TRANSFORM FAULTS

Defining the physical properties of a transform fault is difficult. The only available information is that obtained from the seismicity and from the bathymetry. The petrology of the rocks dredged from fracture zones have also given some clue to the properties of these features but are limited in their value since their exact source is unknown. Gravity data has also provided some insight into the structure of transform faults.

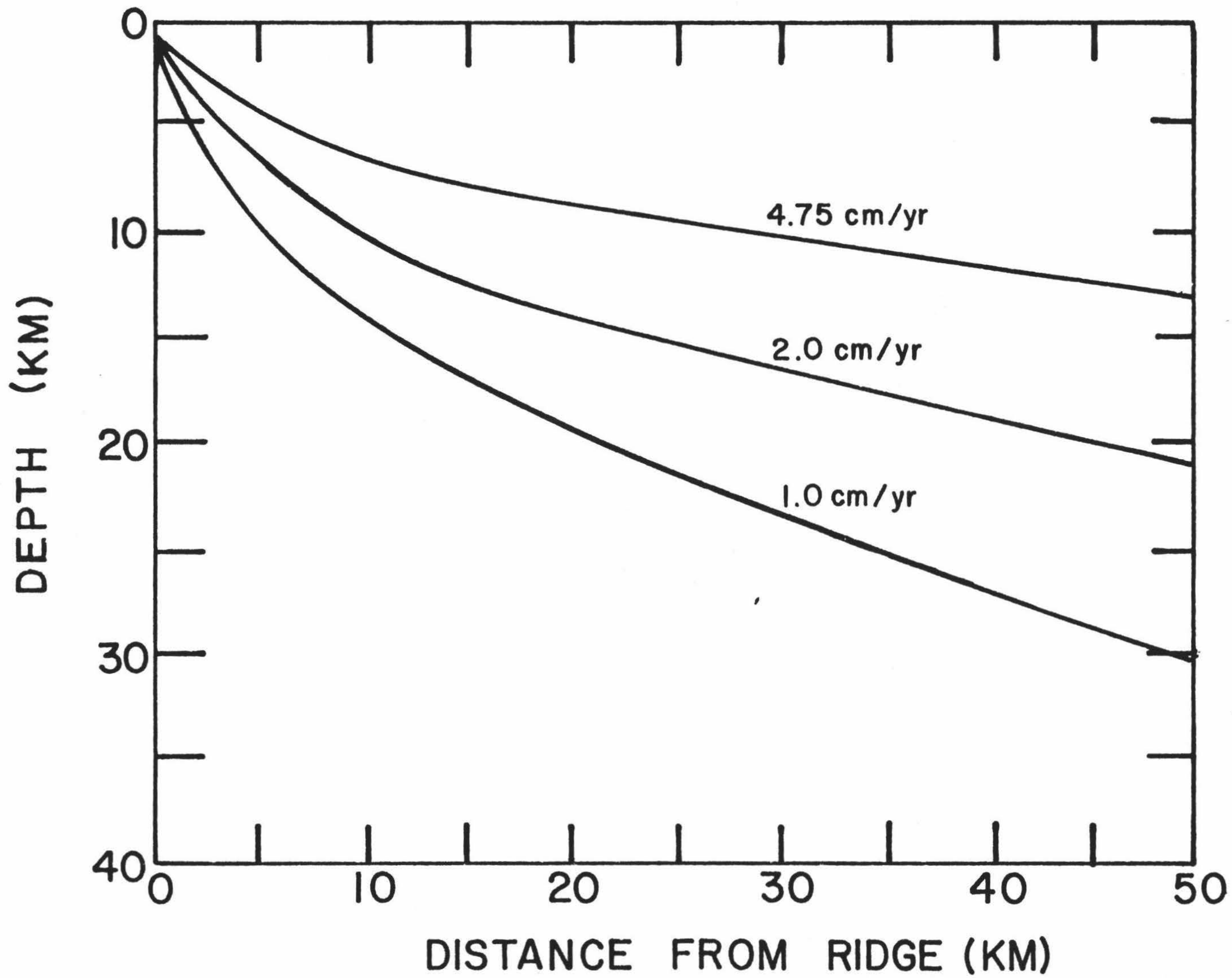


Fig 4a. Isotherms for the solidus (1150°c) calculated for different spreading rates. This is from a flat slab model of thickness, 100 km. with a temperature of 1200°c at the base. Figure from Sleep (1969).

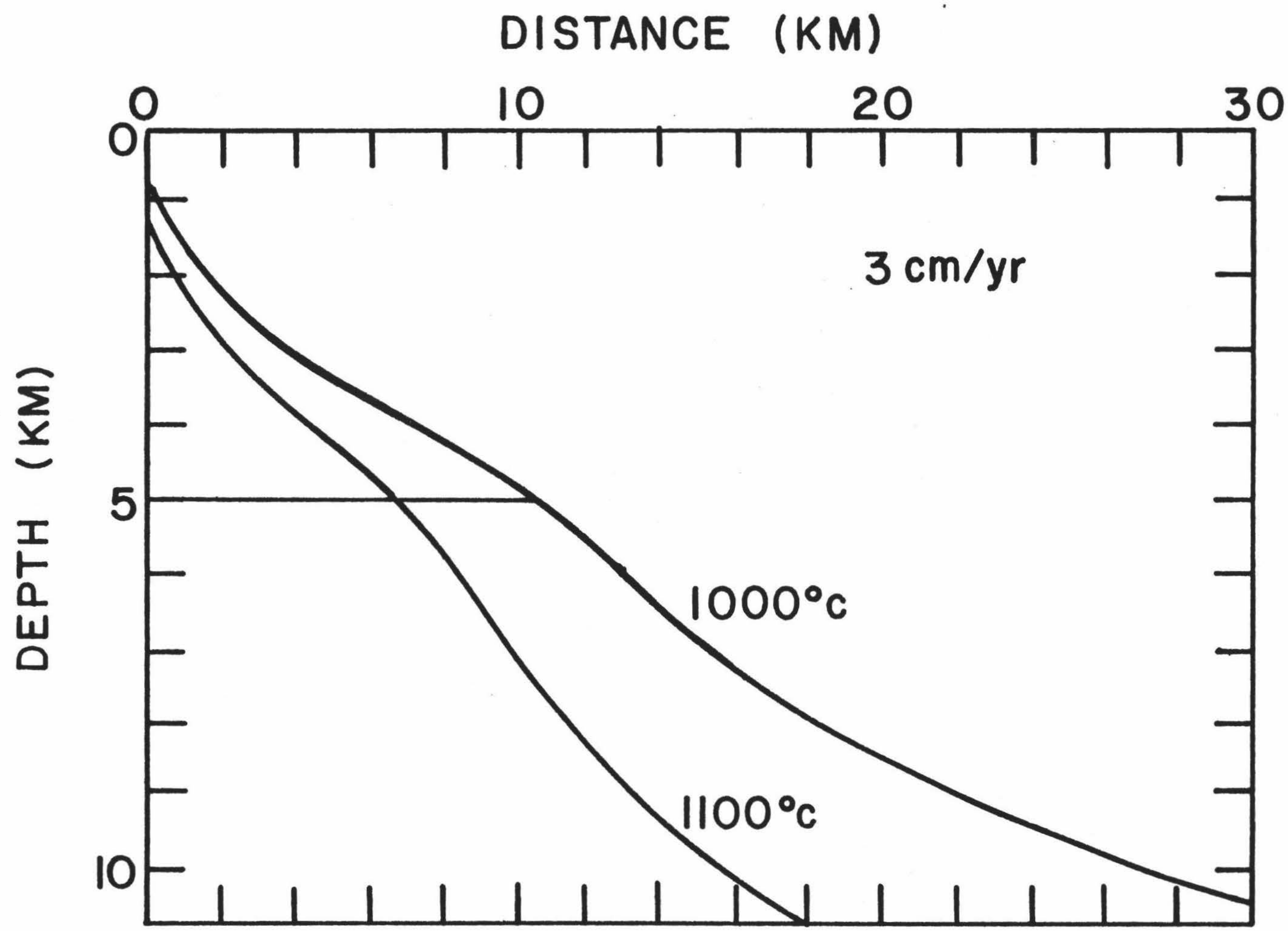


Fig 4b. Position of the magma chamber as proposed by Kusznir and Bott (1976). This was calculated with the effect of the latent heat of fusion included. Base of the chamber is at 5 km. (base of layer 3 in the oceanic lithosphere).

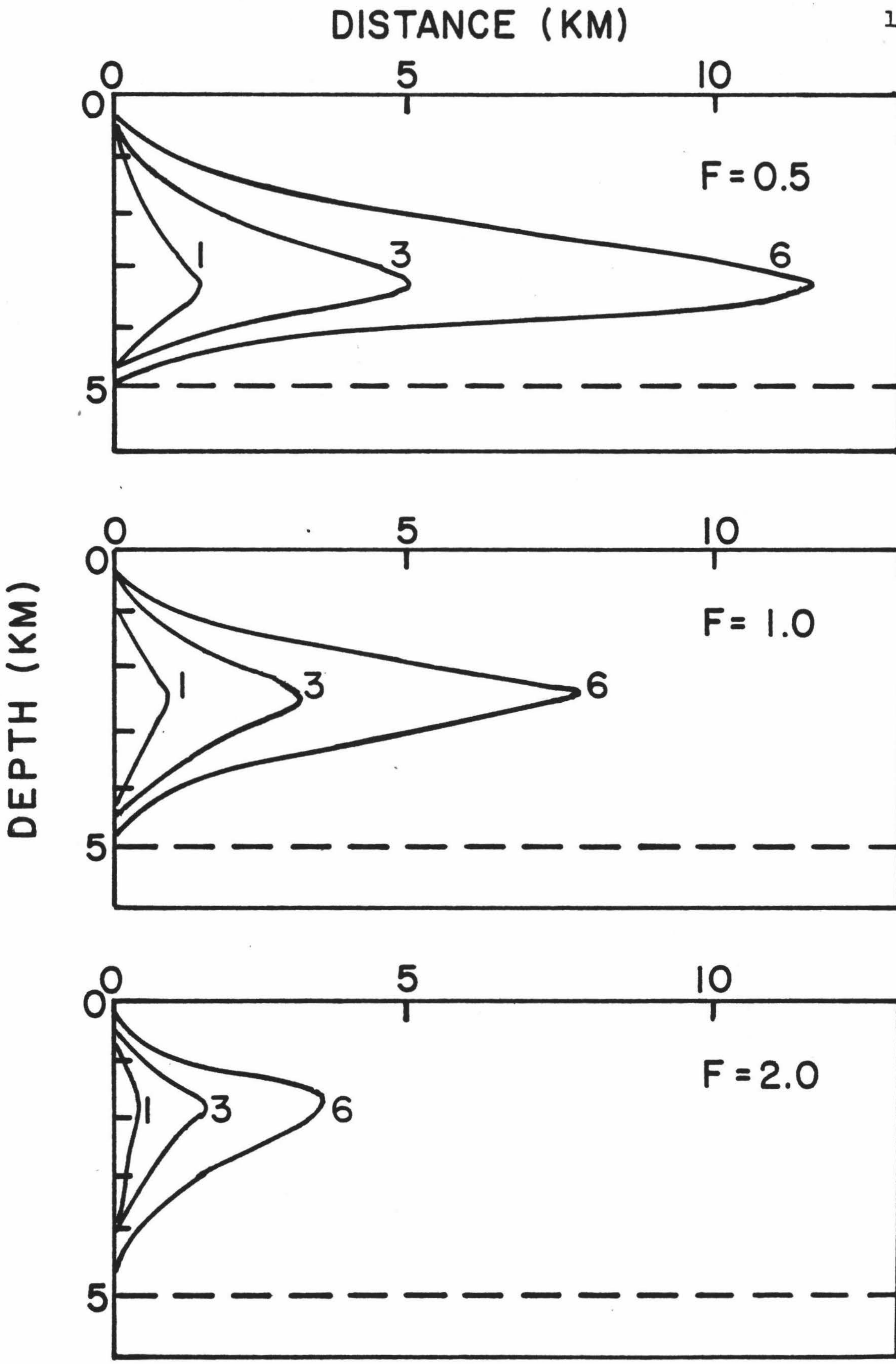


Fig. 5 Position of the magma chamber as based on using the the effect of accretion and crystal settling. The variable F is defined in the text. Number next to curves (1,3,6) are spreading rates in cm/yr. Figure from Kusznir and Bott (1976).

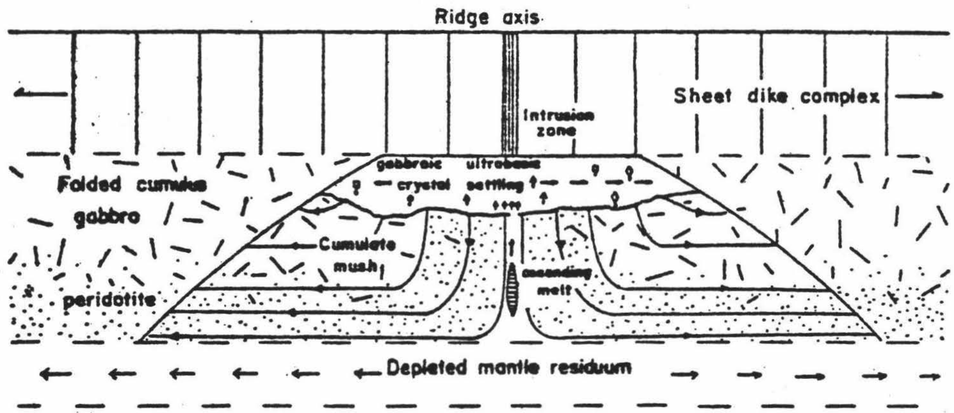


Fig. 6. Schematic diagram showing the hypothetical structure of the magma chamber. Figure is that of Sleep (1975).

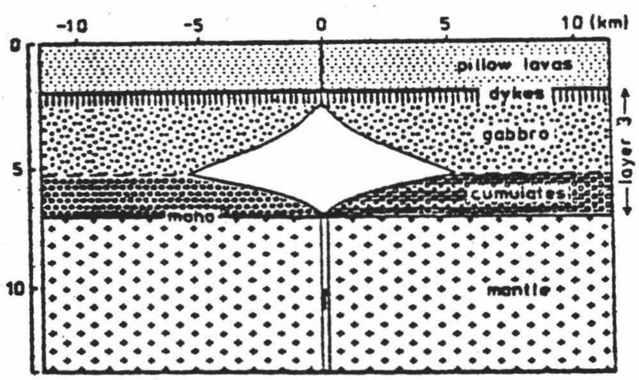


Fig. 7. Structure of the oceanic lithosphere under a mid-ocean ridge. Figure is from Kusznir and Bott (1976).

Brune (1968) in his analysis of seismic moments of various fault zones, determined that the zone of earthquake production lies between the depth of 2-7 km. Due to the relatively shallow depth and small width of the earthquake producing zone, it is expected that no large earthquakes (magnitude greater than 8) will occur on the oceanic transform faults since not enough stress can be built up. This aspect reflects upon the relatively thin nature of the oceanic lithosphere near the transform fault and the high temperature that exists near its base. Recently, a more detailed analysis of the earthquakes that occur on oceanic transform faults has been performed (Burr and Solomon, 1978). In this study it was put forth that the zone of earthquake production can be defined by the area between the 75°c and the 150°c isotherms (a more conservative estimate would be 50-300°c). It is believed that oceanic lithosphere in this temperature range behaves like a brittle substance and can hence accomodate the accumulation of more stress than would a visco-elastic or viscous material.

It has been argued that the mid-ocean ridges tap a source region which is widespread and generally homogeneous in the upper mantle (Kay et al., 1970). However, when large offsets are present in the ridge system, such as in the case with the northeast Pacific (eg. Mendocino fracture zone) then it is expected that some discontinuities may exist (Fig. 8). Furthermore, the rate of spreading should also have some effect since it will change the location of the asthenosphere-lithosphere boundary.

If a mid-ocean ridge does tap a general source region, then such a region should extend under the associated transform faults. This will have some bearing if a leaky transform fault (to be discussed later) exists and can then reach this source area. Problems with this are that the depth to the source material may vary under certain areas and that the petrology of rocks taken form the sea-floor can be derived from a range of different sources, thus making a proof of this idea difficult to attain. Vogt and Johnson (1975) have argued that if a hot spot is situated near a transform fault, then a form of longitudinal flow of asthenospheric material can take place (Fig. 9). This will produce a form of damming effect with the net result being the production of

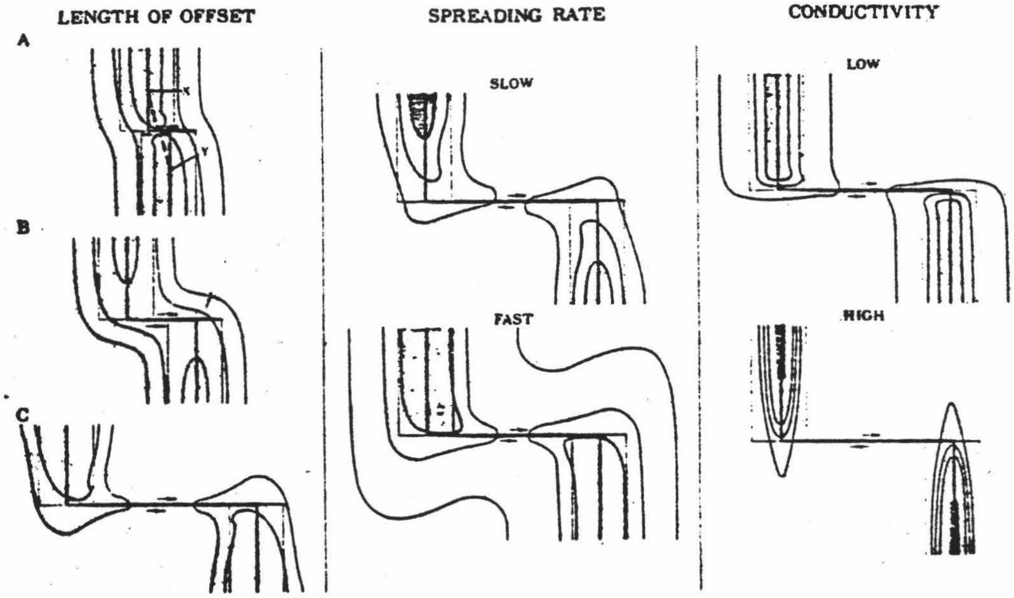


Figure from Vogt et al, 1969

Fig. 8. Depth to a particular isotherm, as well as lithosphere thickness must depend on offset on transform fault, spreading rate, and conductivity. In this purely schematic diagram, the darker shading indicates the thinner lithosphere. For small offsets a new crack *XY* can readily destroy the transform fault because the crack need only break through a short section of thin crust and have a large radius of curvature. The greater the offset, the deeper the "energy trap" of a transform fault. One implication is that transform faults with short offsets may come and go during the history of an ocean basin, whereas the great faults survive. A second implication is that, in general, the higher the spreading rate, the greater the minimum offset and, for a given regional trend, the greater the spacing between fracture zones.

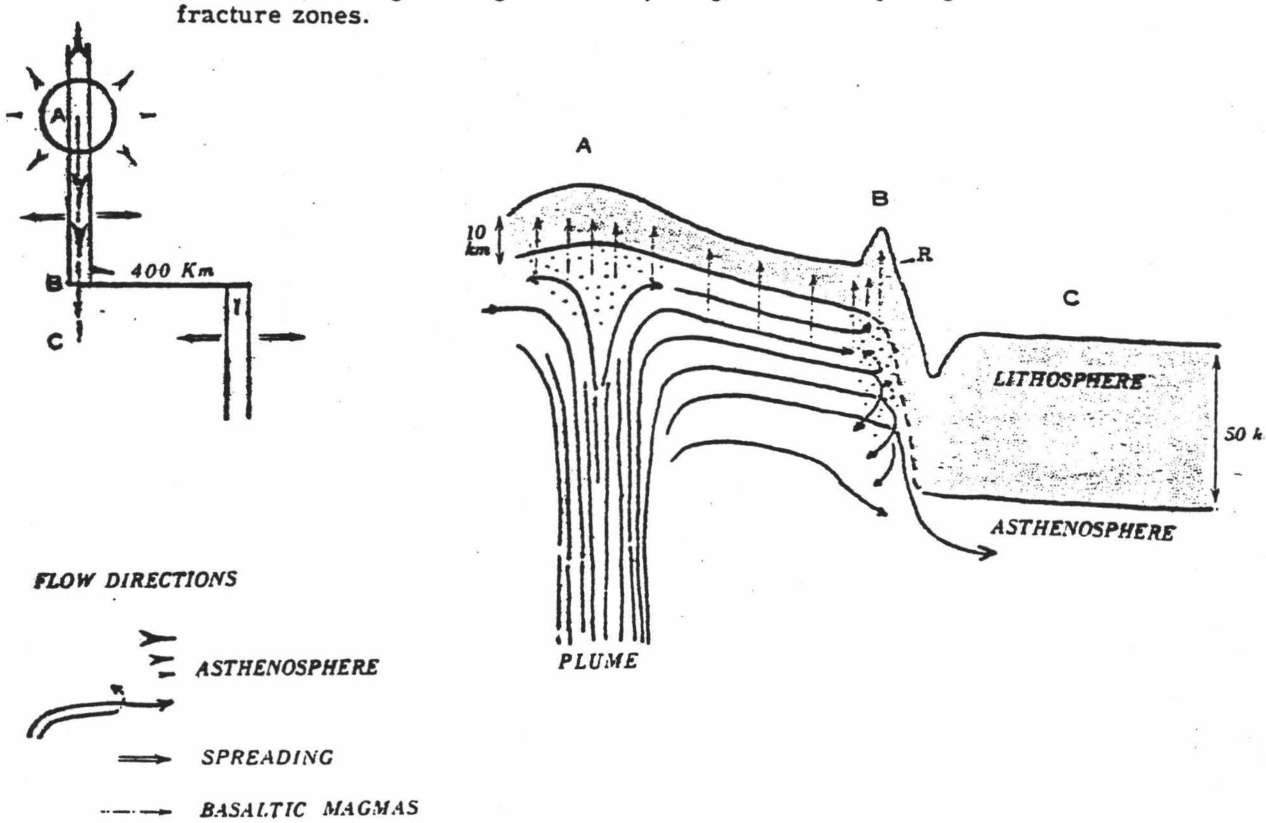


Fig. 9. Illustration of the longitudinal flow of asthenospheric material which can produce aseismic ridges along transform faults. Figure is from Vogt and Johnson (1975).

aseimic ridges. Furthermore, in some cases, a hot spot may not be needed for this damming effect to occur. This has led to a more systematic analysis of these ridges.

Dredgings off some of these aseismic ridges in the Atlantic have shown these ridges to be composed mainly of ultramafic material, predominantly gabbro, peridotite and serpentinite (Bonatti and Honnorez, 1976). It has been proposed that these ridges represent uplifted sections of the lower oceanic crust (layer 3) and therfore provide a window into the lower oceanic crust. Recently, it has been proposed that the mechanism for uplift is related to the thermal contraction of the plates as it moves away from the ridge crest (Turcotte, 1974; Bonatti, 1978). This fracturing of the lithosphere enables water to reach a lower depth and therefore the process of serpentinization will occur in layer 3. This produces sufficient low density material which thereby raises sections of the crust. The validity of calling these ridges, sections of oceanic lithosphere has been questioned by Francheteau et al. (1976) in that the tectonic environment of a transform fault is anomalous when compared to that for the rest of the oceanic lithosphere.

For large enough offsets beween ridges, a considerable thermal and topographical anomaly will exist near each ridge. On one side of the fracture zone and/or transform fault, very young material will be present while on the other side relatively older crust will be present. Louden and Forsyth (1974) analyzed this situation and realized that there should be a gravity edge effect along with thermal conduction across the fracture zone. This effect will decrease as the offset between the ridges decreases and as the age of the lithosphere increases.

Another property of fracture zones can be seen when the plate is subducted. Barazangi and Isacks (1976) in their seismicity study of the Peru-Chile trench have proposed that the plate can tear along pre-existing factures. This is possibly the result of differences in the thickness and density of the oceanic lithosphere on the different sides of a fracture zone. Furthermore, the likelihood of fracturing should increase when the age of the offset across the fracture zone increases and when the age of the subducting lithosphere decreases(ie. the difference in thickness of the plate is greatest when the lithosphere is relatively young).

Thus it is seen that the older the lithosphere, the less the difference across the fracture zone will be. DeLong et al.(1977) have proposed using an idealized, perfect model, that a fracture zone should have both dip slip and strike slipe motion on it. Furthermore, this should continue for a considerable length of time (at least 120 m.y.). This is supported by an apparent bathymetric difference across fracture zones in the Pacific in areas of about this age. The extent at which the fracture zone "heals" is unknown although it is generally assumed the these features should diminish in time.

More pertinent to this paper is a discussion on the relative stability and configuration of a ridge-transform fault system. Initially it was generally assumed that transform faults always formed perpendicular to the strike of the ridges. Lachenbruch andThompson (1972) applied a linear resistance model to explain this configuration. In their model they assumed that the ridge-transform fault system would reallign itself such that the resistance to sea-floor spreading would be minimized. They proposed a S/r_o ratio where S=resistance to slippage along the transform fault and r_o=resistance to spreading at the ridge. The simplified form is

$$S/r_o = 2 \tan\theta \qquad (8)$$

where θ is the angular difference from a perpendicular system. This concept has been questioned recently by Atwater and Macdonald (1977). They present evidence that in a slow spreading environment, oblique spreading and oblique ridge systems are quite common. However, in a fast spreading environment (v greater than 3.0 cm/yr), nearly perpendicular configurations are found. Therefore, a very long, normally alligned transform fault in a fast spreading environment would have a S/r_o ratio which is very small. On the otherhand, a very short, oblique transform fault in a slow spreading environment would have a large S/r_o ratio. This is intuitively wrong as some factor is needed to take into account the length of the transform fault. Turcotte (1974) also points out that an application of a linear model to a nonlinear situation may not be valid.

LEAKY TRANSFORM FAULTS

Menard and Atwater (1968, 1969) were the first to propose that a change in the direction of sea-floor spreading may result in the crustal extension of a transform fault, thereby resulting in the formation of what they termed a "leaky transform fault". The simplified model is shown in Fig. 10. In this example, a change in the direction of sea-floor spreading (angle θ) has occurred between time t_o and t_1. No absolute measurement of the magnitude of either the time, the angle, nor an explanation for why such a change occurs will be given at this time. We are presently interested only in the simplest case. It can be seen from Fig. 10 that the lower transform fault has realligned itself parallel to the direction of sea-floor spreading while the upper one has not. Therefore, if this configuration is maintained, then the upper transform fault must take on a component of spreading (Fig. 11), along with a component of strike slip motion. Furthermore, the magnitude of these components are

$$V_n=V_r \sin\theta$$
$$V_s=V_r \cos\theta \qquad (9)$$

where V_n = normal component of spreading

V_s = strike slip component of spreading

V_r = spreading rate (half rate) at the ridge

θ = change in the direction of spreading (angular change)

For expected values of θ (less than 30°)

$$V_n< V_s< V_r \qquad (10)$$

If this configuration is maintained for some length of time, then the sequence in Fig. 12 is expected. Eventually, for reasons to be discussed later, it is expected that the leaky transform fault will be eliminated. One idealized case calls for the presence of a propagating rift (Fig. 13 b,c). The other possibility is through the formation of another spreading center between the two existing ridges (Fig. 13 d).

One simplified way of modelling a leaky transform fault would be to break it up into two separate mechanisms, one being the strike slip motion of a typical transform fault and the second being the component

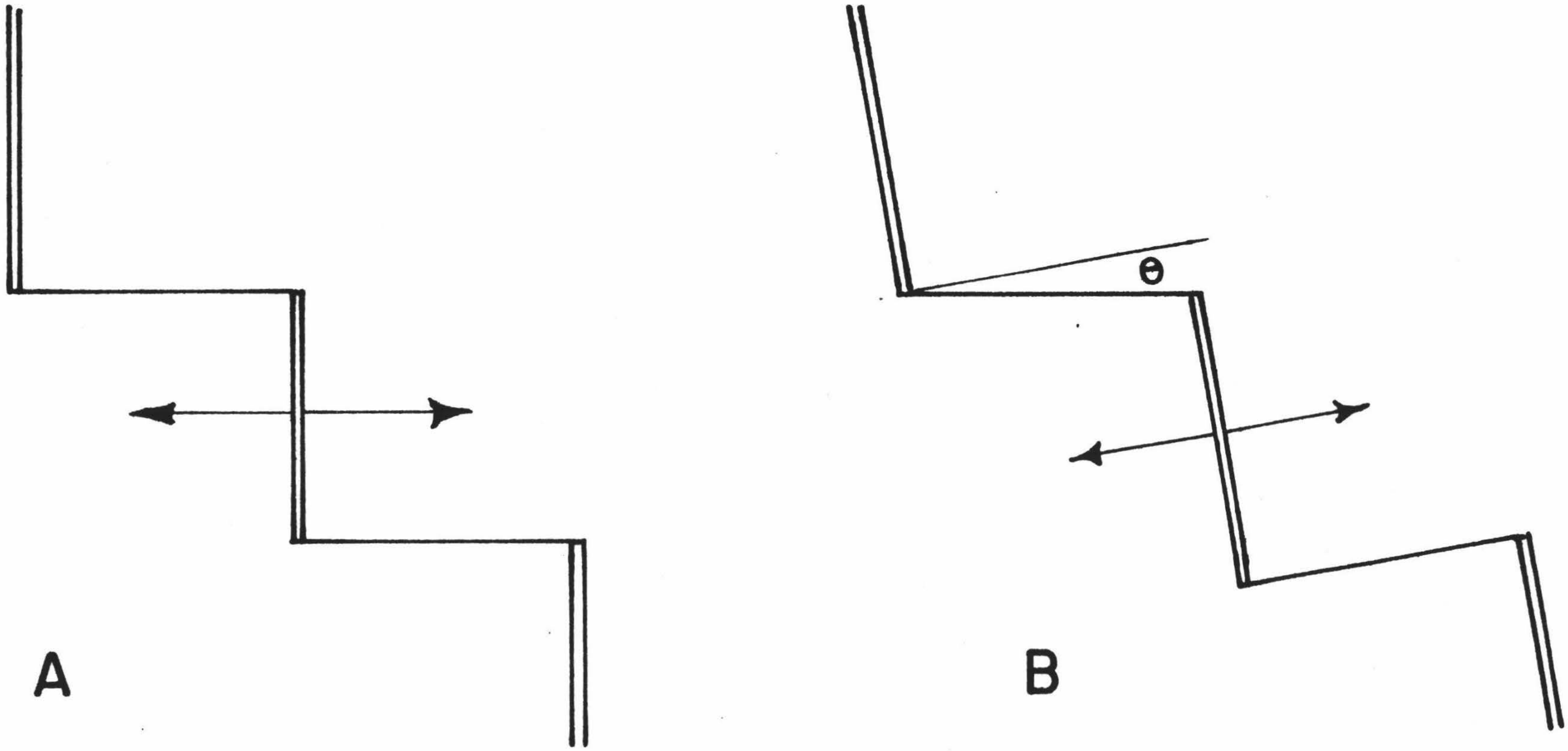


Fig. 10. a) At time, t_o, prior to a change in spreading direction, the ridge and transform fault are perpenddcular.

b) Change in spreading direction of angle θ at time, t_1. Bottom transform fault has realligned itself while the top one is still in its original configuration.

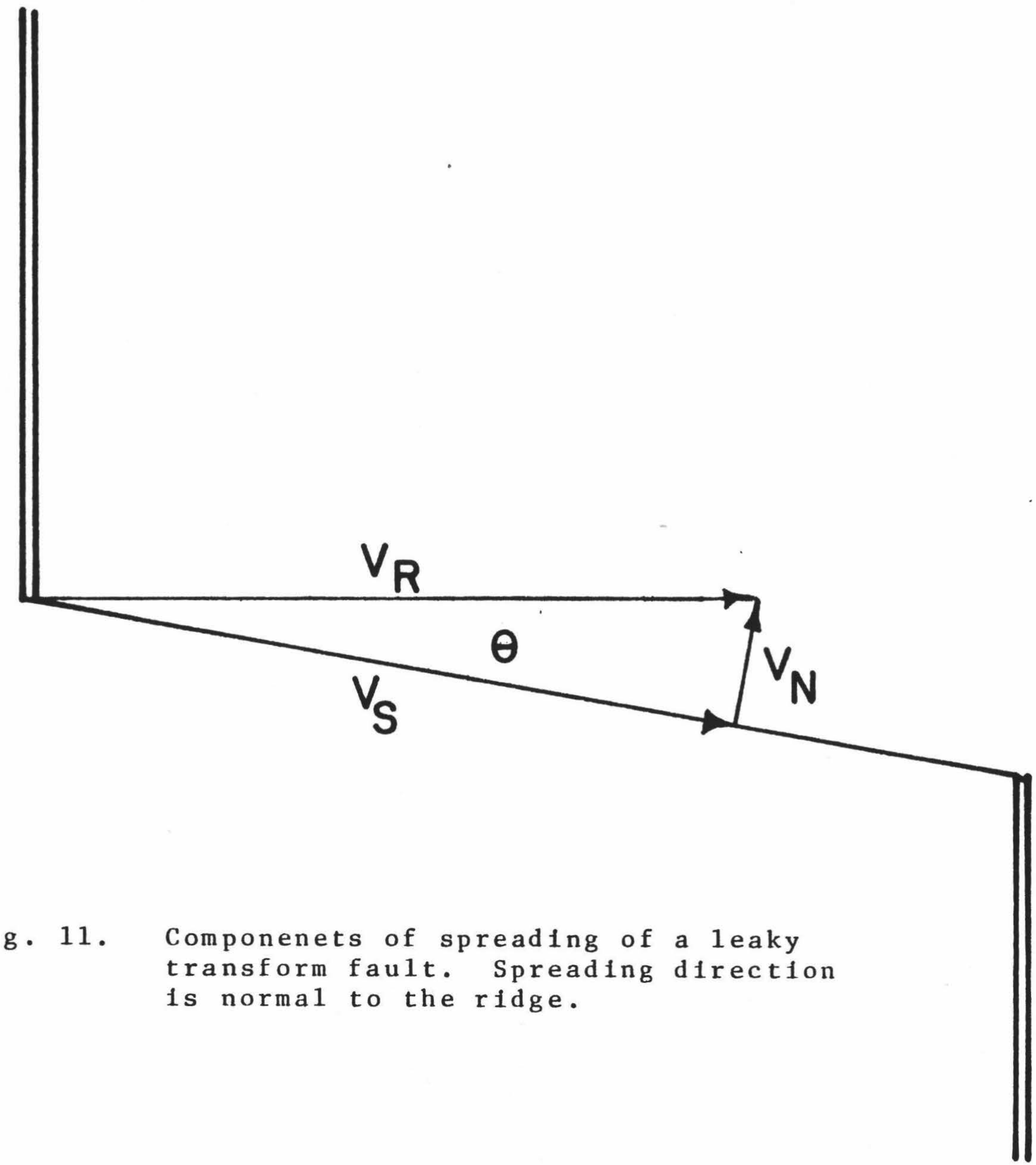


Fig. 11. Componenets of spreading of a leaky transform fault. Spreading direction is normal to the ridge.

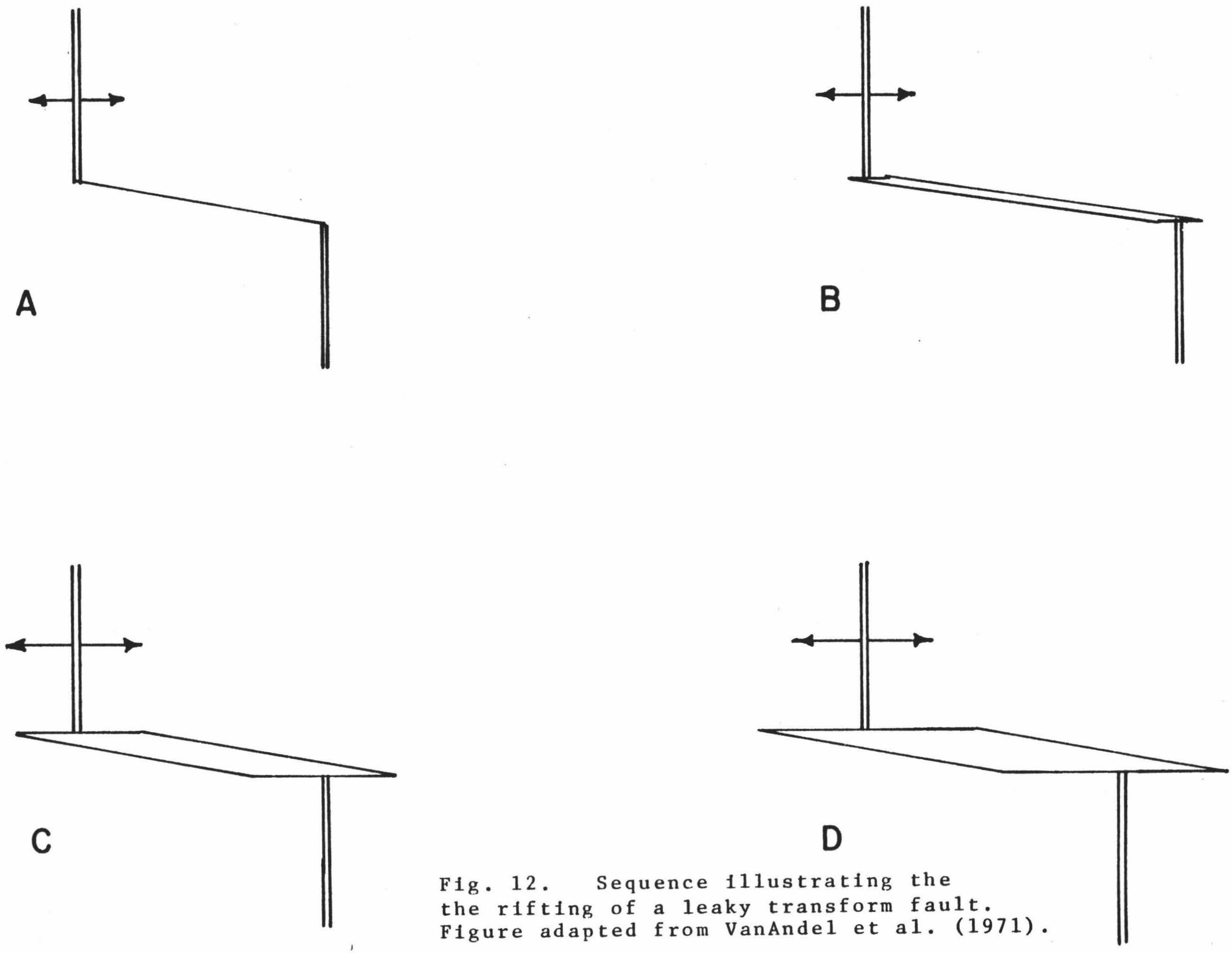


Fig. 12. Sequence illustrating the the rifting of a leaky transform fault. Figure adapted from VanAndel et al. (1971).

of spreading. If one views the spreading motion as being representative of being a mid-ocean ridge, then a thermal model can be set up. By using the new coordinate system as shown in Fig. 14, the heat flow equation now becomes

$$\rho\, C_p\, [\, \frac{\partial T}{\partial t} + \vec{v}\, \vec{\nabla} T\,] = \text{div}\, [\, \kappa \vec{\nabla} T\,] + H \tag{11}$$

where the velocity, $\vec{v}$, is now

$$\vec{v} = \hat{i}\, v \cos\theta = \hat{j}\, v \sin\theta \tag{12}$$

If we further assume that $H=0$ and that the transform fault is long enough and that the point under consideration is far enough away from the ridge, then the steady state equation is

$$\rho C_p\, [\, v\cos\theta \frac{\partial T}{\partial x} + v\sin\theta \frac{\partial T}{\partial y}\,] = \kappa [\frac{\partial^2 T}{\partial x^2} + \frac{\partial^2 T}{\partial y^2} + \frac{\partial^2 T}{\partial z^2}] \tag{13}$$

However, if we neglect the effect of the strike slip motion, then this equation reduces to the generalized heat flow equation of McKenzie (1967)

$$\rho C_p\, v \sin\theta \frac{\partial T}{\partial y} = \kappa [\, \frac{\partial^2 T}{\partial y^2} + \frac{\partial^2 T}{\partial z^2}\,] \tag{14}$$

with the velocity now equal to $v\sin\theta$.

By making the previous assumptions and simplifications, we are reducing a leaky transform fault simply into the case of a slow spreading ridge with no topography. But before any conclusions can be drawn from this model, some of the implications should be reviewed.

By neglecting the terms for strike slip motion, this model is only valid if there is very little friction between the two plates sliding past each other. This is quite different from than the case for a typical transform fault. Therefore, either the coefficient of friction and/or the normal compressive force must be small in order for the frictional forces to be small (frictional force = normal force*coefficient of friction). Intuitively, this is a reasonable assumption since the leaky transform fault is supposedly rifting apart (tensile stress) and secondly because of the possibility of molten material (dike intrusions) between the two plates. Furthermore, the oceanic lithosphere in the vicinity of a leaky transform fault is relatively young and therefore still subject to plastic rheological behavior.

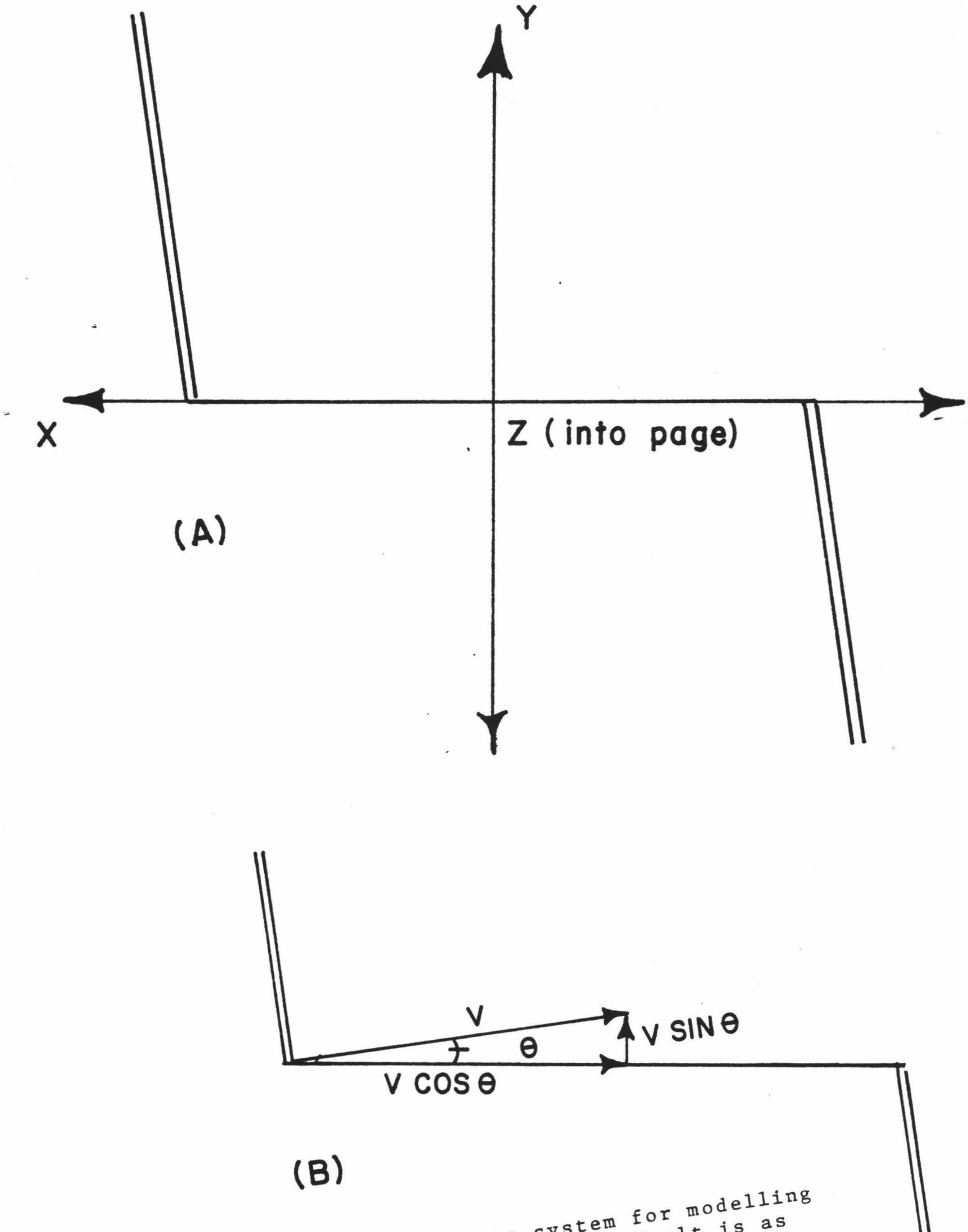


Fig. 14. a) Coordinate system for modelling a leaky transform fault is as shown.
b) The new spreading rate components are illustrated.

If such intrusions exist, then it should modify the thermal structure of the transform fault. This in turn should affect the earthquke producing zone by making it smaller since additional heat is added to the system (Burr and Solomon, 1978). Therefore, a possible test of the leaky transform fault hypothesis is to do a detailed seismicity study comparing a typical transform fault and a possible leaky transform fault. Such a study would require a large data set along with a good knowledge of the seismicity of a typical transform fault.

The assumption that the effects of the nearby mid-ocean ridge is small is questionable. Louden and Forsyth (1976) have shown that there is thermal conduction across fracture zones due to the differential ages on opposite sides. This in turn may help the mid-ocean ridge to propagate, thereby eliminating the leaky transform. Furthermore, the regions close to one of the ridges is expected to be highly affected by the heat conduction of the spreading center. However, since we have placed the coordinate system at the midpoint between the two mid-ocean ridges, then the age and thermal characteristics at this point are the same across the transform fault. This model would then break down at some point which is much closer to one ridge than the other. However, the thermal structure of the mid-point is still affected by the regional field of the ridges and therefore this model may only work for long transform faults.

Nonetheless, if this model holds true, then several predictions can be made. Since the normal component of rifting is proportional to the sine of the angular change in dirction, then this value is expected to be rather small. In the case for the MAR (V_r=1.5 cm/yr) with an arbitrary value of θ of about 15°, then it follows that

$$\begin{aligned} V_n &= V_r = (1.5\ \text{cm/yr})*(0.26) \\ V_n &= 0.4\ \text{cm/yr} \end{aligned} \tag{15}$$

This value is below the minimum spreading rate required by the previously mentioned thermal models and therefore if the leaky transform fault can be described by this model, then very little ocean-bottom volcanism is expected. This is the case for the Atlantic where the transform fault ridges have been shown to be composed of ultramafic material and is not basaltic sub-surface volcanic products (Bonatti and Honorrez, 1976). Probably the best example is the Vema fracuture zone along the MAR (around 10°N) which has been proposed to be the result of rifting without the pressence of ocean-bottom volcanism (VanAndel et al., 1971).

For the case of the East Pacific Rise (V_r=8 cm/yr), a completely different situation is found to exist. Once again assuming a change in direction of 15°, then it follows that

$$V_n = V_r \sin\theta = (8 \text{ cm/yr}) * (0.26)$$
$$V_n = 2 \text{cm/yr} \tag{16}$$

This rate is even greater than the spreading rate of the MAR and therefore ocean-floor volcanism is highly probable. It has been proposed that the Sala y Gomez ridge is an example of a leaky transform fault (Principal, 1974; Clark and Dymond, 1977) with the change in the direction of spreading brought about by the plate reallignment that is currently going on (formation of the Easter Plate?). Naturally, the real world does not behave completely like this idealized model and imperfections abound. However, the fact remains that despite the presence of numerous possible leaky transform faults in the Atlantic, no appreciable amount of volcanism associated with "leaking" is found to exist. Furthermore, it has been proposed by Menard and Atwater (1969) that the Moonless Mountains (located just south of the Murray fracture zone) represents the remains of a leaky transform fault.

Finally, another working model of a leaky transform fault is to represent it as a series of short ridges and transform faults (Fig. 15). It can be seen from this figure that the total amount of transform fault is not affected by the number of short ridges used. Consequently, the total length of these short ridges is also a constant. This model is therefore favorable primarily because there is no need to produce some new form of plate boundary. The familiar features of a ridge and a transform fault are all that is required.

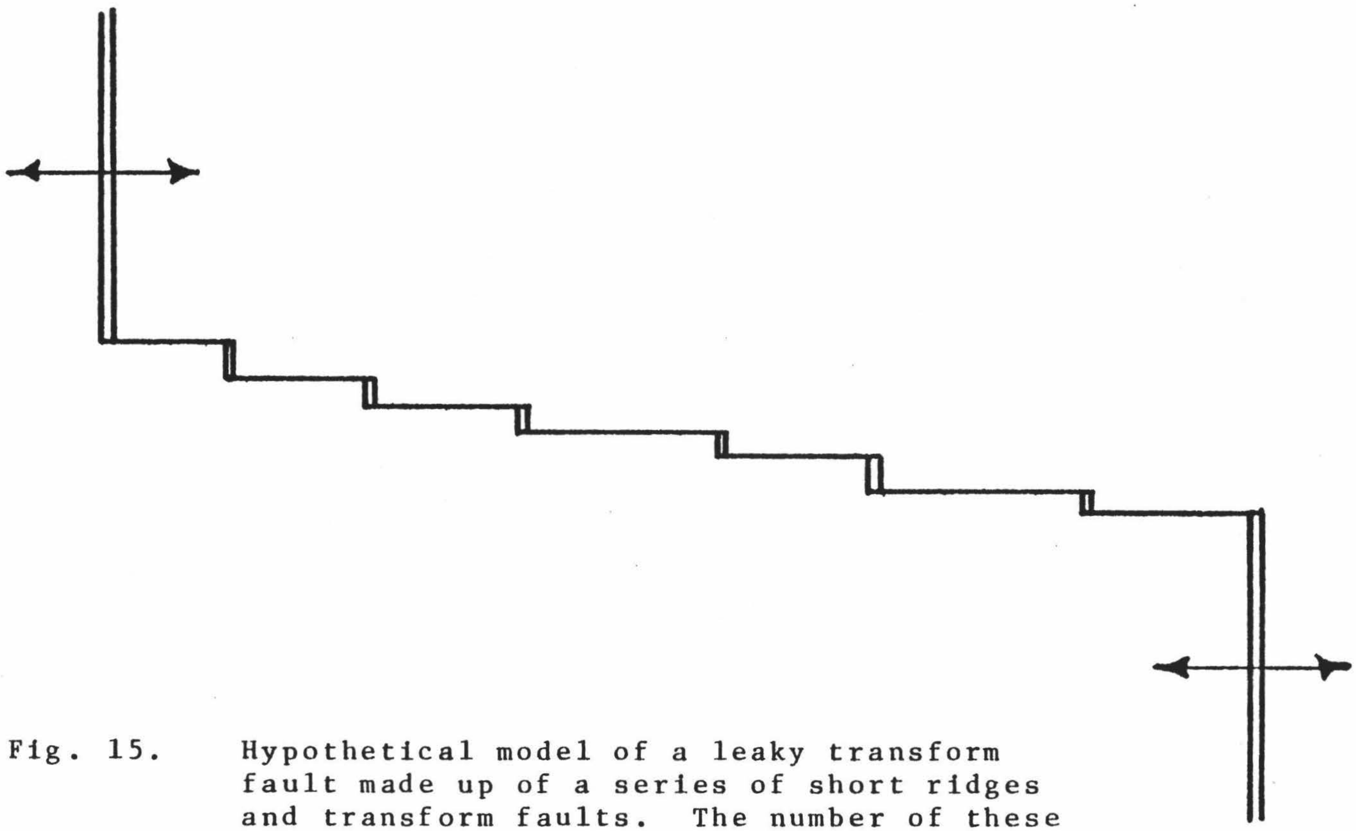


Fig. 15. Hypothetical model of a leaky transform fault made up of a series of short ridges and transform faults. The number of these ridges is arbitrary.

Before we can begin to analyze the stability of a mid-ocean ridge system, it may prove useful to review very briefly some of the current ideas concerning the driving mechanisms of plate motion.

DRIVING MECHANISMS

In defining the driving mechanism of plate motion, one must always keep in mind the limitations involved. Purely mechanical models are still beyond the current state of the art and therefore, most attempts have been made utilizing the contraints of the observed pattern of plate motion or intraplate stress deduced from mid-plate earthquakes. Presently, the two leading models utilize the pull of the downgoing oceanic lithosphere in a subduction zone (Forsyth and Uyeda, 1975; Chapple and Tullis, 1977; Harper, 1975) or secondly, the potential energy developed by the upwarping of the plates at mid-ocean ridges (Solomon et al., 1975). In the first case, the downgoing slab is of a higher density than that of the surrounding asthenosphere due to the contrasting temperature differences. This would therefore have the effect of negative buoyancy, thereby pulling the plate down into the trench. Furthermore, seismic evidence currently indicate that the upper portion of the sinking slab is in tension while the lower half is in compression (Isacks and Molnar, 1969). The compressive stress is probably due to the dissipation of the negative buoyancy effect as the slab reaches a zone of resistance in the mantle.

The second mechanism for moving the plates is caused by the known mechanisms that occur at the mid-ocean ridges. In this case, buoyant mantle material rises to a significant height thus forming the characteristic topography of a ridge system. This in effect supplies potential energy to the system and can therefore be viewed as a process where the oceanic lithosphere is sliding down off the ridges. If this is the case, then compressive earthquakes are expected to occur within the plate, this being in disagreement with the effect predicted by the downgoing slab model in which tensile earthquakes are predicted. In reality, it is the net effect of both compressive and tensile stress along with the localized stress field which determines the type of earthquake that one finds in a particular area.

Another point to be made concerning the mid-ocean ridges is that energy is required to rift the two plates apart (Lachenbruch, 1976). In other words, these features represent passive structures once the effect of the elevation of the ridges is taken away. Initially it was thought that the injection of magma into the ridges produced a pushing apart effect. However, it is now believed that since the molten material eventually accretes to the conduit walls, energy is needed to pull them apart (Lachenbruch, 1976).

Other forces have been mentioned as being the primary mechanism for driving the plates. Initially it was though that convection cells in the mantle were responsible for plate motion (Hess, 1962). However, this hypothesis is believed to be incorrect (Richardson et al., 1976) based upon intraplate stress. Furthermore, it has been proposed that a form of mantle drag is operating between the lithosphere and the asthenosphere which is acting as a resistive force (Forsyth and Uyeda, 1975). Whether this drag resistance is stronger under the continents than under oceanic lithosphere is still debatable although most indications are that it is greater under the continents (Chapple and Tullis, 1977).

Another mechanism which had been proposed is that of hot spots or mantle plumes (Morgan, 1971). In this model, the lithosphere is essentially sliding off the upraised plate at a hot spot. However, this model is not compatible with either intraplate stress or plate motions.

There has been one more force proposed which requires mention. This is the force due to the downgoing slab in trenches pulling the plate with the trench . This is known as trench suction (Forsyth and Uyeda, 1975). It has been recently proposed that this is the only force needed to drive the plates (Shoemaker, 1978) although supporting evidence is lacking.

Besides the driving forces, there are also resistive forces present. These include transform fault resistance, asthenospheric drag, slab resistance in the trenches, colliding resistance and additional resistance under the continents (Forsyth and Uyeda, 1975; Chapple and Tullis, 1977). Of these forces, the only ones that are worth mentioning are slab resistance in the trenches and asthenospheric drag (mantle drag). The latter one has already been mentioned and will not be repeated here. The other one, slab resistance, is the force that a downgoing slab feels as it reaches

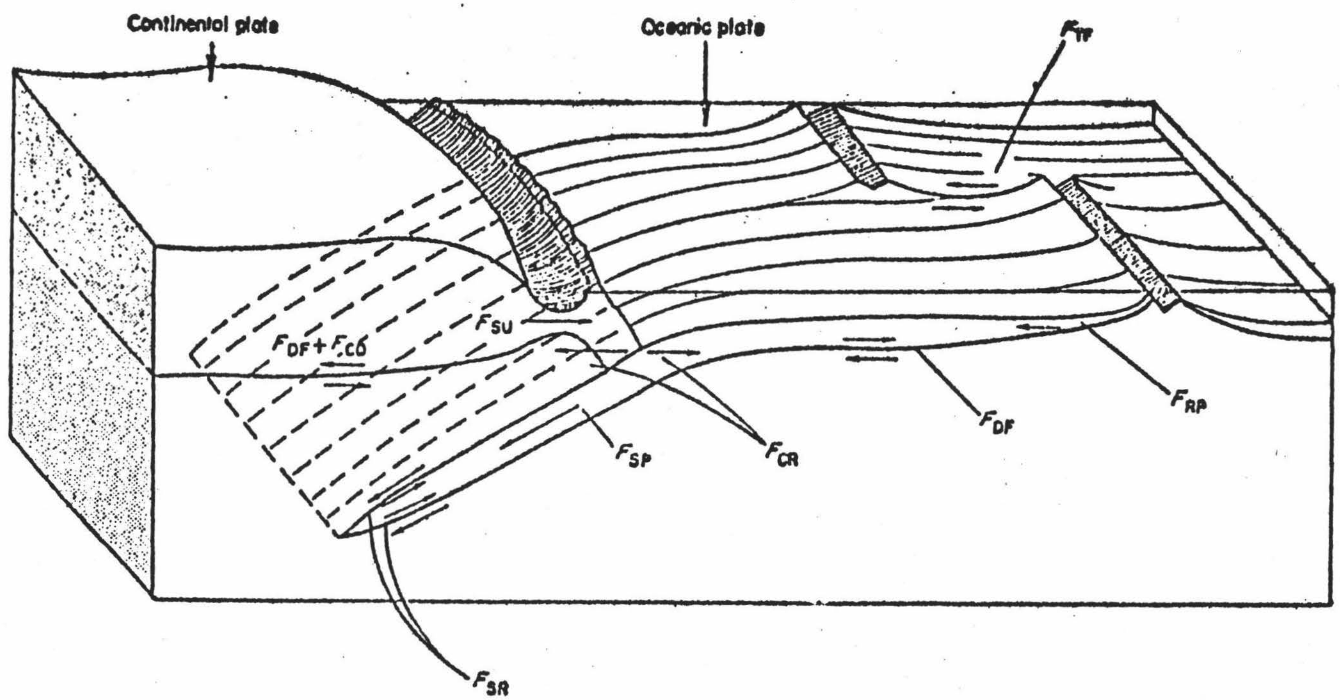


Fig. 16. Schematic diagram illustrating the possible forces that are acting on the plates. These include slab pull in the trenches (F_{SP}), ridge push (F_{RP}), trench suction (F_{SU}), colliding resistance (F_{CR}), transform fault resistance (F_{TF}), mantle drag (F_{DF}), slab resistance in the trenches (F_{SR}), and continental drag (extra mantle drag under the continents (F_{CO}).

sufficient depth in the mantle. This force is considered to be quite large although not as large as that of slab pull. Therefore, the net effect of the downgoing slab in a trench will be a driving force. A composite diagram of these forces is given in Fig. 16.

With this, we essentially finish the first half of this paper. We can now look at some of the various ways in which a ridge system can deform.

DEFORMATION OF A MID-OCEAN RIDGE SYSTEM WITH EMPHASIS ON THE NORTHEAST PACIFIC

Despite the relative stability of mid-ocean ridge systems, it is known that they deform and undergo substantial reallignment. Generally, it has been assumed that such deformation is the result of changes in the regional stress field acting upon the ridge system. The two general changes are in the direction of spreading and in the rate of spreading. But there are also other types of deformation that can occur.

Ridge jumps have been identified to be a common feature on the ocean floor. The most prominant one proposed to have occurred in the region under investigation took place in the area between the Murray and the Molokai fracture zones (Harrison and Sclater, 1972; Menard and Atwater, 1968). The distance of the jump was about 530 km and is believed to have occurred around 37 myBP (using the LaBreque et al., 1977 time scale) (Fig. 17). In this case, such a jump appears to be quite conceivable since this eliminated the large offset in the ridge system. The new ridge could then tap a source region that was continuous under the East Pacific Rise.

Usually ridge jumps on the order of the previously mentioned distance can produce some interesting problems. If the model for the upper mantle below a mid-ocean ridge presented by Forsyth (1977) is correct, then the jump will break through a section of the plate that is about 9 m.y. old. If the isotherms are correct, then the dry solidus would be found at a depth of around 40 km. This is very close to the position of the adiabat. If the adiabat is crossed (refer back to Fig. 3), then substantial melt would not immediately form. Therefore, some time lag is expected to be present and the formation of a magma chamber will be delayed.

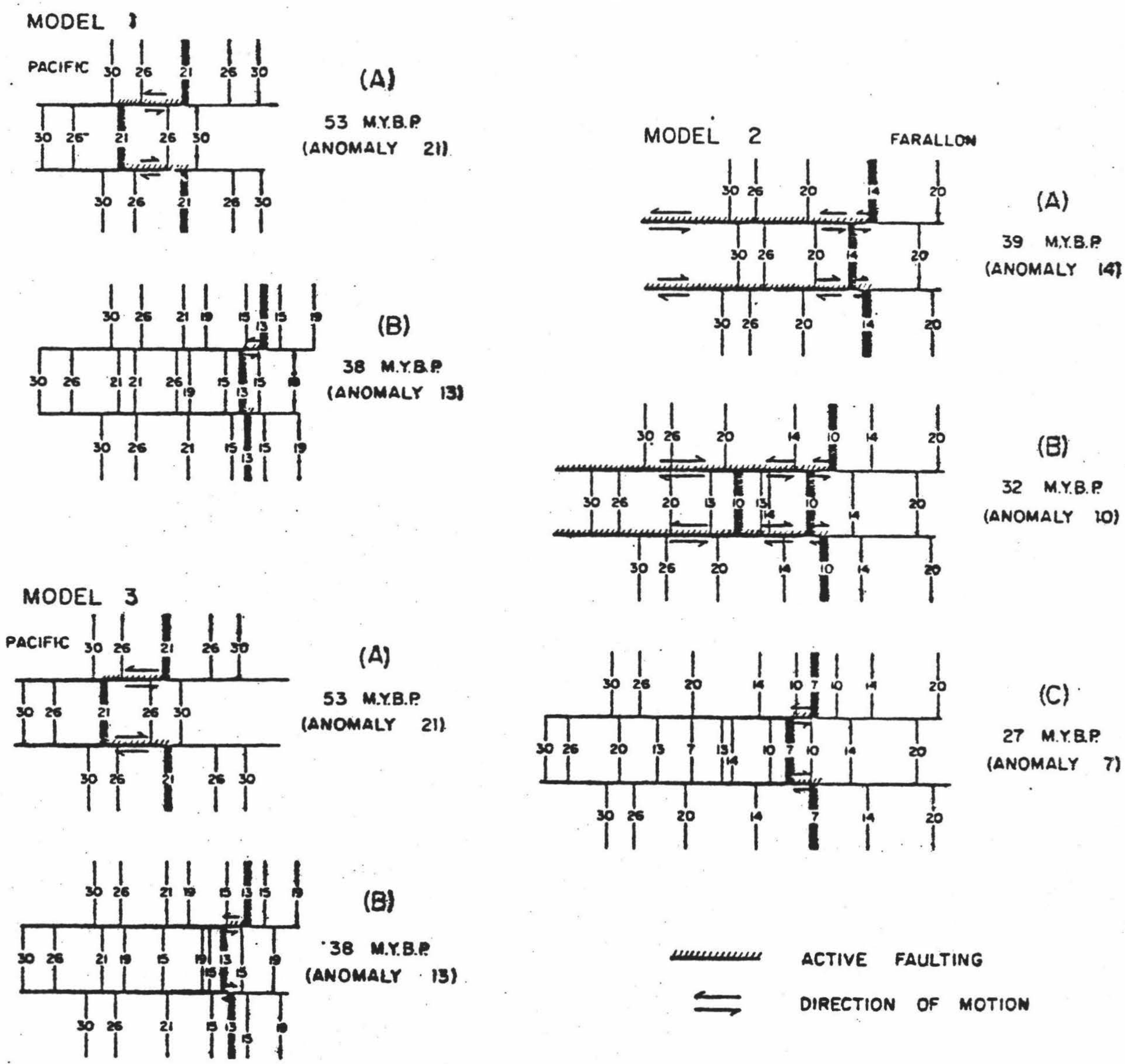


Fig. 17. Three models to explain the disturbed zone. Model 1 is from Menard and Atwater (1968). Model 2 is that of Malahoff and Handschumacher (1971). This model has two ridges active for a period of time. The final model is that of Harrison and Sclater (1972). All the times are calculated using the Heirtzler et al., (1968) time scale and can hence be updated by using the newer LaBreque et al., (1977) time scale. Figure is from Harrison and Sclater (1972).

Another series of ridge jumps has been proposed to have occurred north of the Mendocino fracture zone (Shih and Molnar, 1975). This occurred over a longer period of time (anomalies 13-5E, 36-18 myBP based upon the LaBreque et al. time scale) and is believed to be the cause of the elimination of the Surveyor fracure zone. More importantly is the concept that the ridge may have propagated (Fig. 18).

Such properties as ridge jumps and ridge propagation (propagating rift) have been identified on a finer scale by Hey (1977; Hey and Vogt, 1977) in the Galapagos region. It has also been proposed that such a propagating rift is responsible for the formation of the "zed" pattern of Menard and Atwater (1968) as in the case of the Juan de Fuca area (Hey, 1977).

The model for a propagating rift has some interesting features. One such model has only one of the ridges propagating with the other ridge being passive at this point. As shown in Fig. 19b, the transform fault has realligned itself such that it connects the two ends of the ridges. However, this configuration is known to be unstable since it is rarely seen in the real world. Therefore, for stability to be attained, the other ridge must shorten and the transform fault must then reallign itself accordingly (Fig. 19c). If this process were to continue for a period of time, then a sequence similar to that shown in Fig. 19 is envisioned.

The key point is that the second ridge does not shorten until sufficient stress is built up by the unstable transform fault configuration. In other words, the propagating rift produces a region of compression around the transform fault which then in turn produces the resulting reallignment. This model is expected to work best in a fast spreading environment. This is because such deviations as oblique spreading and oblique transform faults are usually not seen (Atwater and Macdonald, 1977).

Another interesting point is to allow both ridges to propagate but not necessarily at the same time. Then the following sequence is possible (Fig. 20). This would produce a sheared zone of some finite width. Assuming that the ridge can only propagate until the stress on the transform fault is too great, then it can be determined how wide this

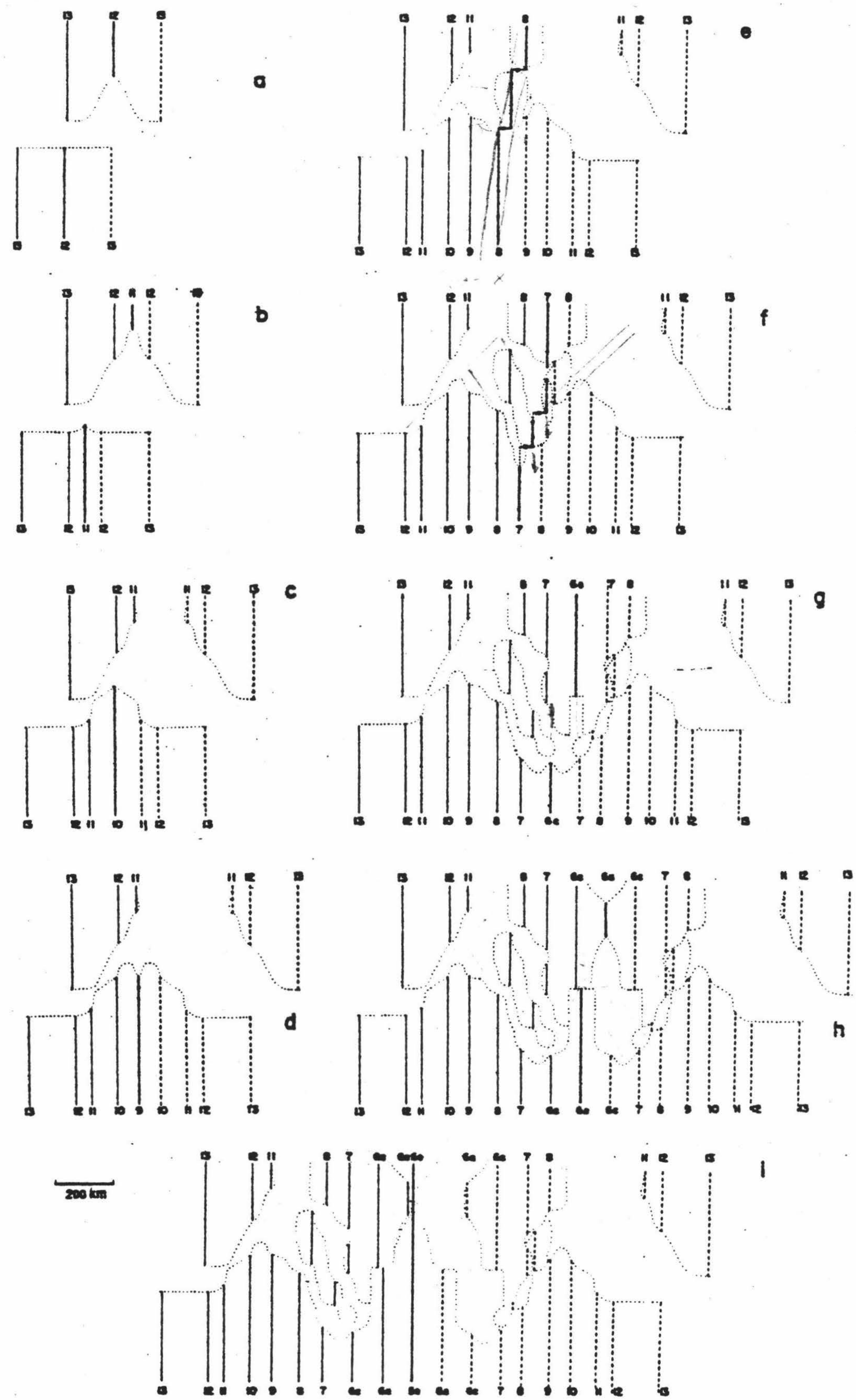


Fig. 18. Schematic sequence showing configuration of the Pacific-Farallon spreading center at the times of anomalies 12,11,10,9,8,7,6C,6A,and 5E. Dotted lines enclose regions where crustal accretion has been interpreted to be symmetric Figure is from Shih and Molnar (1975).

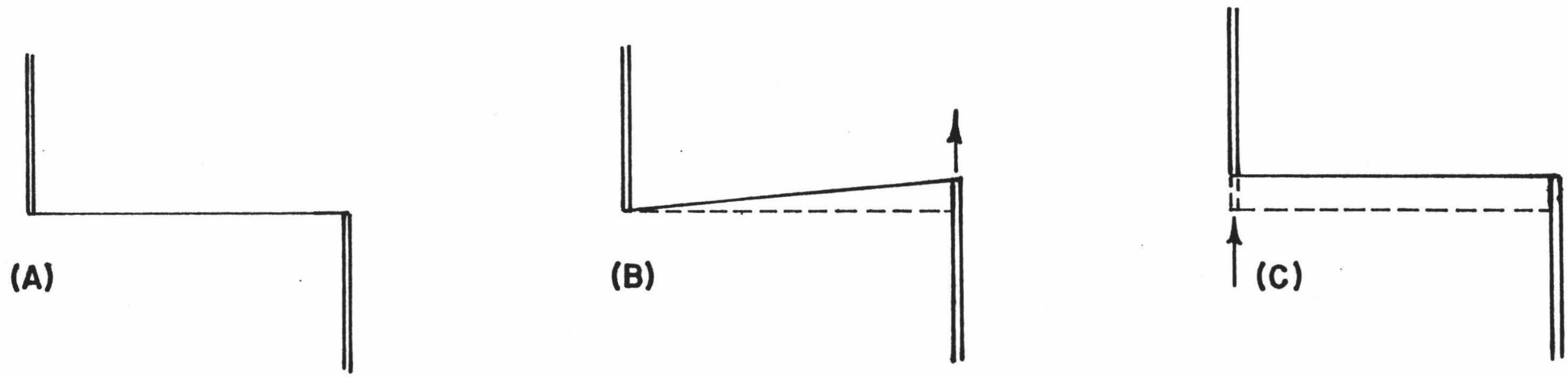
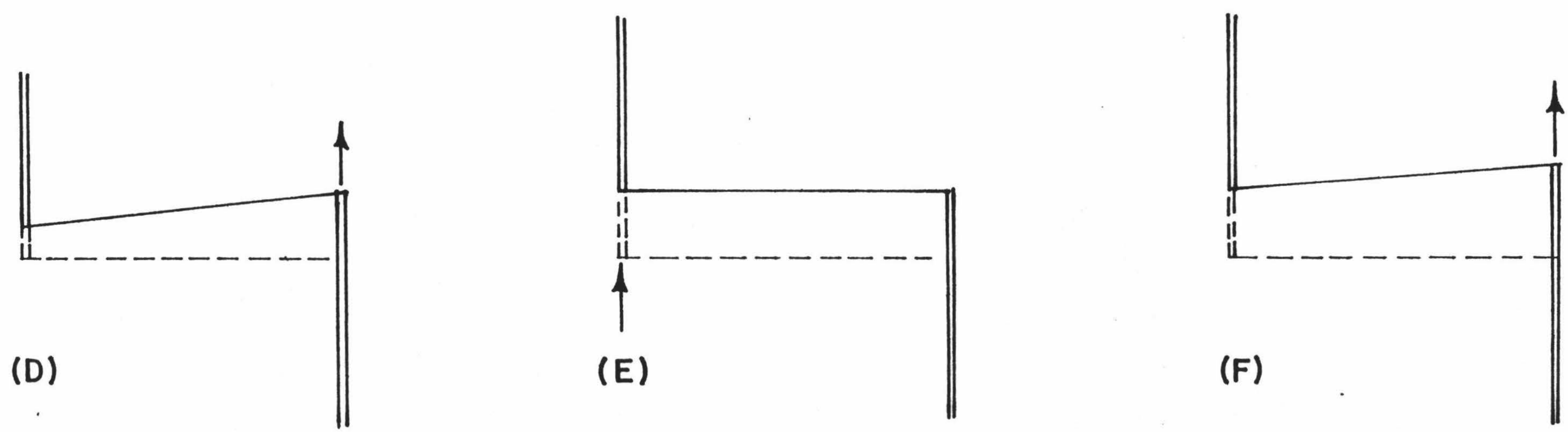


Fig. 19. Propagating rift model of Hey (1977). Time sequence is from (a) to (f). Dotted lines show the position of the inital configuration of the ridge and transform fault.



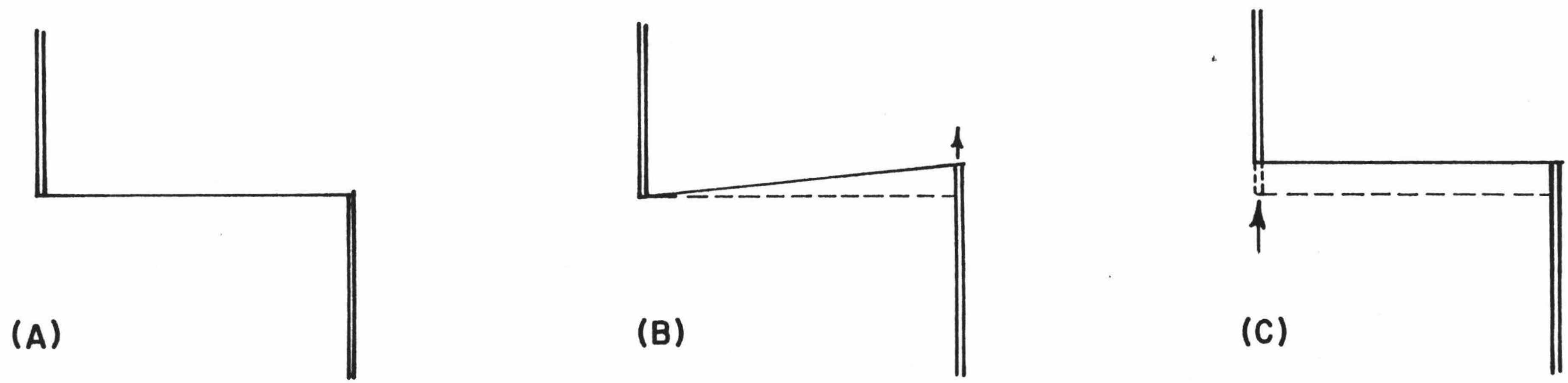
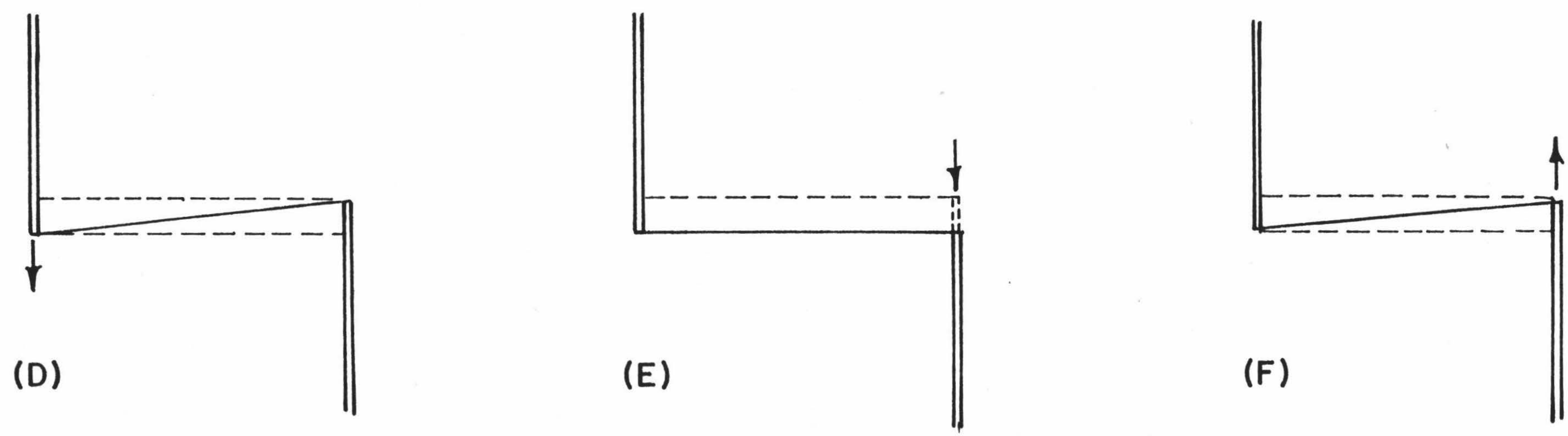


Fig. 20. Propagating rift model which could possibly explain the rifted nature of some fracture zones. Time sequence is from (a) to (f) but is also repeating (a to e).



sheared zone should be. For example, assume that the maximum angle of obliqueness is 5° and that the length of the transform fault is 300 km. Then the width, W, is

$$W = L \tan\theta$$
$$W \cong 26 \text{ km.} \tag{17}$$

(L = length of the transform fault and θ = angle of obliqueness) This model could explain some of the apparent rifted nature of some of the fracture zones in the northeast Pacific (eg. Murray fracture zone) and may prove to be an alternate mechanism instead of a leaky transform fault.

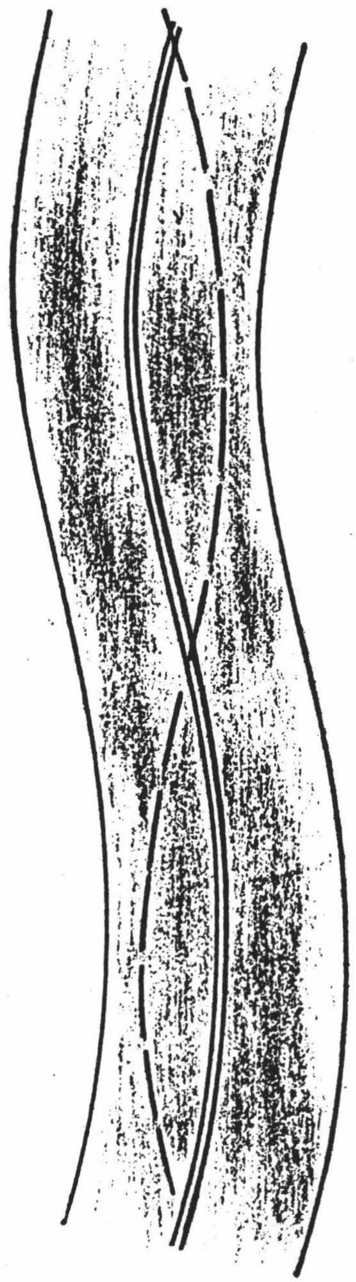
Asymmetric spreading is also a common feature and is on a fine scale usually the case rather than the exception. Stein et al. (1977) have presented evidence that when asymmetric spreading occurs, the ridge migrates in the direction of the trailing plate. This model is based upon the application of some simple fluid dynamic properties and is supported by observations in the real world.

Another characteristic of ridge systems is a type of sinusoidal pattern that they produce in the spreading pattern (Blakely, 1975). This is best illustrated in figure 21. The scale of this ridge migration can vary. Such a pattern has been identified in the region north of the Mendocino fracture zone and along the EPR off the Peru-Chile trench (Rea, 1976).

Up to this point, no mechanical model has been presented to explain these deviations. Recently, Fujita and Sleep (1978, in press) through the use of the finite element method, have analyzed the effect that changes in the orientation of the stress field have on a simple ridge-transform fault system. Their results show that the concentration of stress may lead to reorientation of the ridge through what they proposed to be preferential dike propagation. Furthermore, they also show that when a large offset in the ridge system exists, then the stress will localize in a sense that would shorten the offset.

CHANGES IN THE REGIONAL STRESS FIELD

Changes in the regional stress field can be brought about by a number of different factors. As mentioned before, the subduction of the oceanic lithosphere would transmit tensional stress across the entire plate. Fujita and Sleep (1978, in press) have shown that changes in the regional stress field can have a significant effect on the configuration of the ridge.



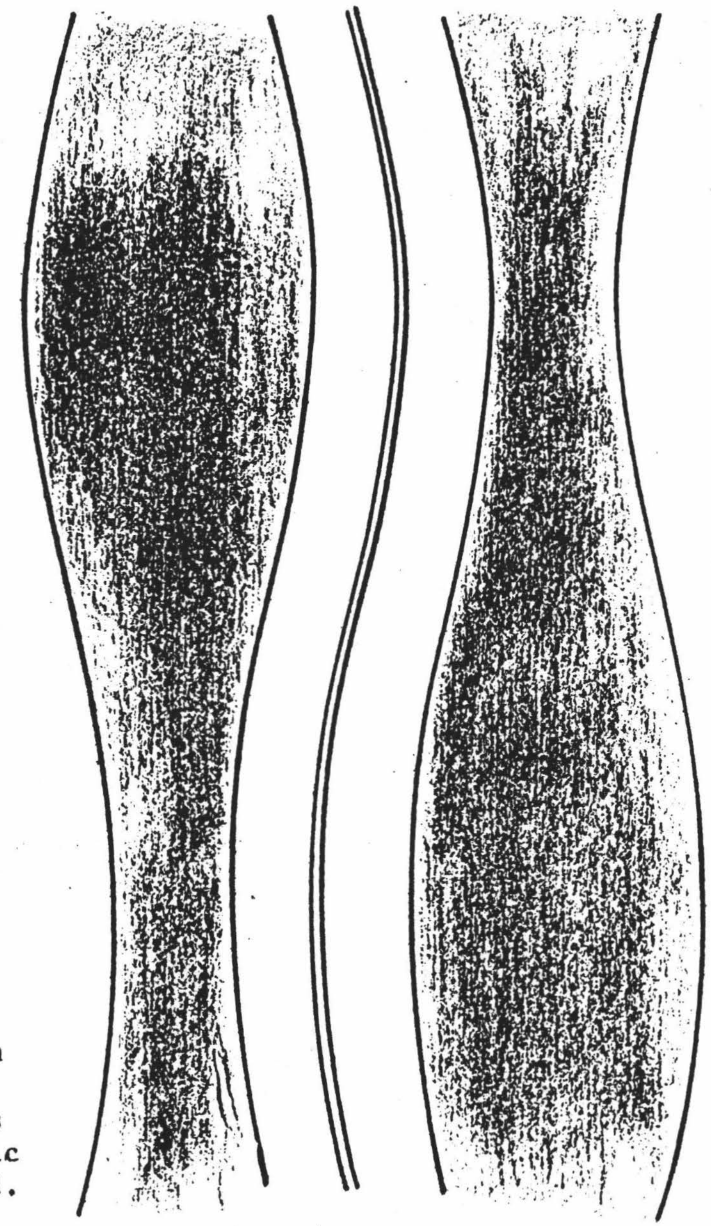


Fig. 21. Sinusoidal pattern produced by a migrating ridge segment. Shaded areas indicate a different magnetic polarity than the non-shaded. Figure adapted form that of Blakely (1975).

In the case of the EPR and the North American trench (NA trench), the configuration of the trench with respect to a section of the mid-ocean ridge must have changed during the onset of ridge subduction. As can be seen from Fig. 22, the obliqueness of subduction (as illustrated by a hypothetical ridge system and trench similar to that of the EPR and NA trench) is not constant over the entire length of the trench. If the two plates are allowed to move in opposite directions, then the expected stress field along sections of the ridge system will change accordingly. Furthermore, if the fracture zones act as stress guides by isolating sections of the plate, then ridge reallignment or jumps could occur only within the affected sections when they are in line with the anomalous section of the trench.

The primary difficulty with such a concept is the determination of the magnitude of the change in the stress field. Since the oceanic lithosphere thickens with age, it is believed that the effect will be minimal when the ridge system is far from the trench (ie. the subducting portion of the plate is relatively old) and the fracture zones had enough time to heal. However, when the ridge approaches a trench and therefore the subducting plate is younger, then the possibility for reallignment and deformation increases.

A simple-minded way of thinking of this is through the use of the relationship $\sigma=\mu\varepsilon$. In this case, σ= stress, μ=some elastic parameter and ε=strain. If σ is considered to be a constant, then as the ridge approaches a trench, μ should decrease as the leading plate is relatively still young (and therefore relatively still hot). This then implies that the amount of strain should increase. This is naturally a gross oversimplification, but the general concepts should hold true.

RIDGE SUBDUCTION AND THE TECTONIC HISTORY OF THE NORTHEAST PACIFIC

A prominant feature of the EPR prior to ridge subduction was the anomalous large ridge offset along the Mendocino fracture zone of about 1000km. based upon magnetic lineation offset. The difference in the age and subsequent thickness of the oceanic lithosphere on the opposite sides of the fracture zone is relatively large as there was approximately

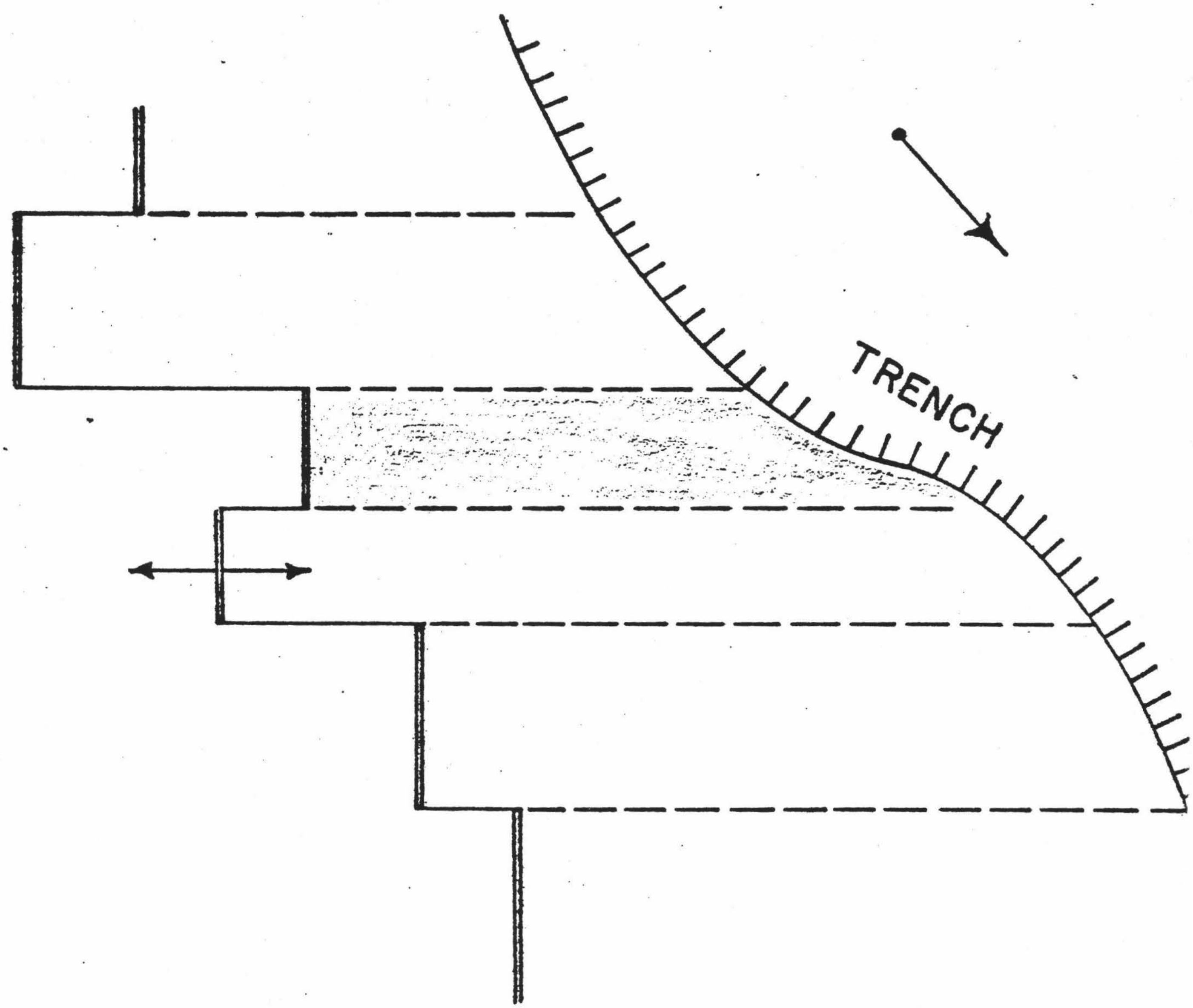
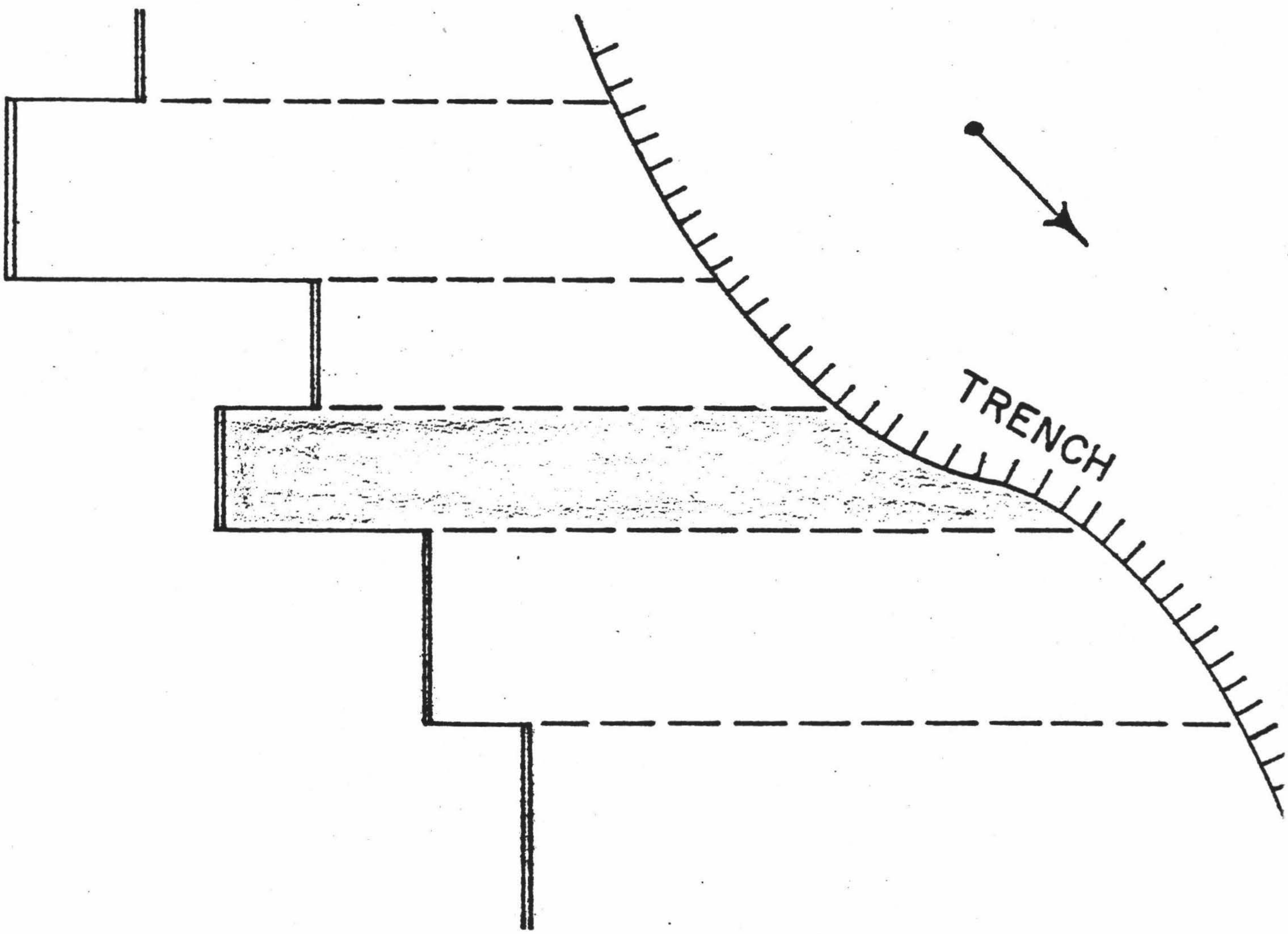


Fig. 22. Diagrams showing how sections of the mid-ocean ridge can be in a different stress field than the rest of the ridge. In this case, the area that is shaded is the one being affected. If the plates are allowed to move in opposite directions, then the following sequence is envisioned.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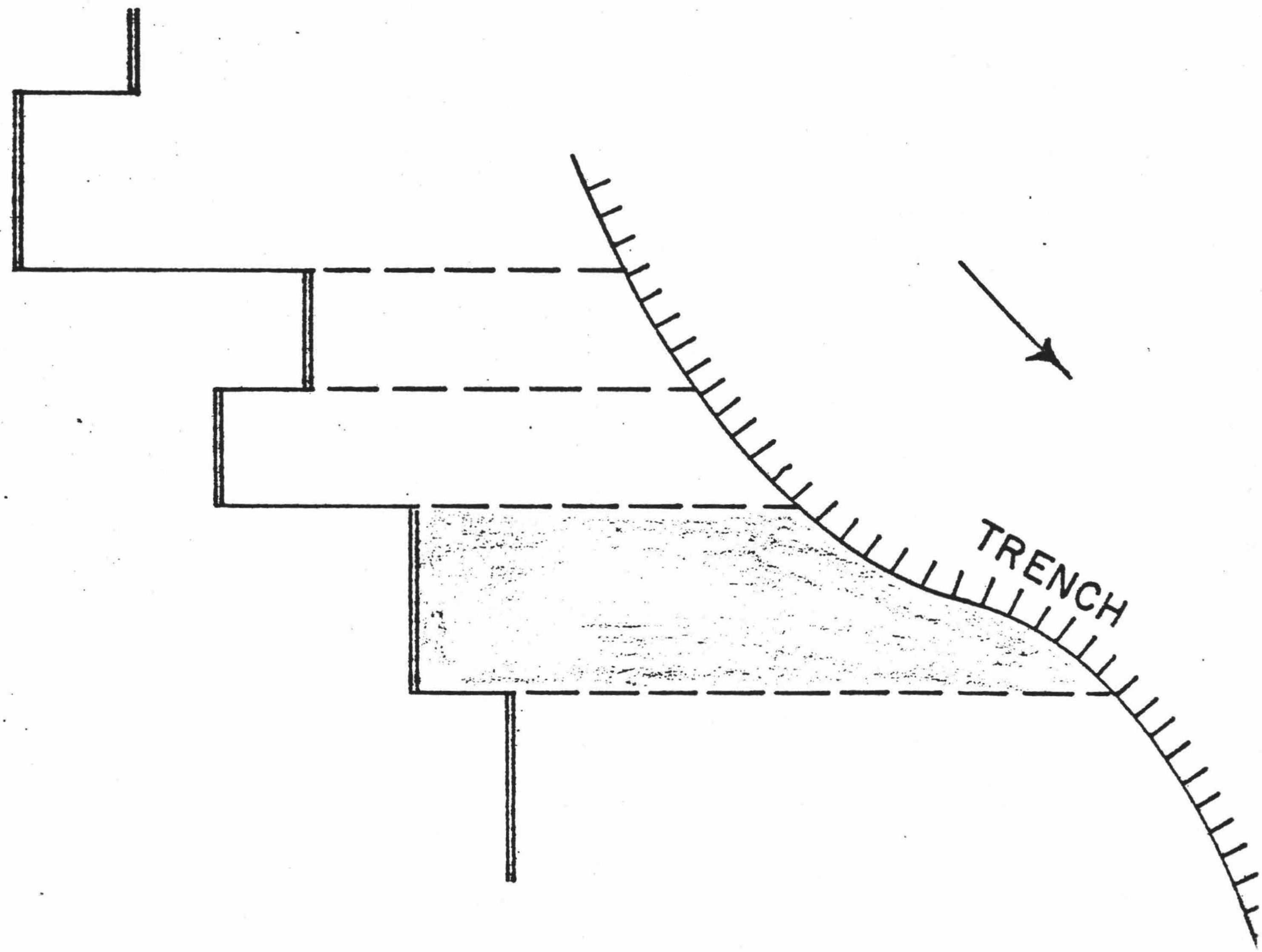
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a 30 m.y. age contrast. Therefore, it is expected that the lithosphere north of the Mendocino fracture zone should have been more stable than the much thinner material to the south for a similar distance away from the trench.

Another significant factor is the characteristics of the region between the Molokai and Murray fracture zones. According to Harrison and Sclater (1972), this area was the site of a substantial ridge jump of about 500 km. Therefore, when the ridge approached the trench, the downgoing slab would have been slightly thicker than that implied from the new configuration. From a qualitative standpoint, this may have some effect upon the timing of spreading rate or directional changes.

Therefore, the downgoing lithosphere to the north of the Mendocino fracture zone and to the south of the Murray fracture zone would have been thicker than the section in between. This implies that the effect of slab pull in the middle section should have been smaller and therefore the stress field acting upon the EPR would not have been constant along its length. Furthermore, this effect becomes more significant the closer the ridge is to the trench. Or in other words, the chances for intraplate deformation, ridge propagation and possible breakup increase.

PREDICTED EFFECT ON THE DRIVING MECHANISM

The mechanics of pre-ridge subduction is poorly known. However, based upon the previous descriptions on the ridge-transform fault system, some speculative concepts can be presented.

As a mid-ocean ridge approaches a trench, progressively younger lithospheric material gets subducted. Due to the relatively younger age, the descending slab is expected to be thinner and hotter than for a normal downgoing plate. The temperature and density contrast between the subducted oceanic lithosphere and surrounding asthenosphere will not be as large and therefore the effect of slab pull in terms of the driving mechanism will decrease. Furthermore, due to the increase in the relative buoyancy of the subducted slab, the dip at which the plate descends will also decrease (Uyeda and Miyashiro, 1974). This should have some effect on the type of volcanism that is occurring behind the trench (DeLong et al., 1978).

Therefore, if slab pull were the only driving mechanism of plate motion (with all other forces being either resistive or very small), then the spreading rate for the EPR opposite the North American trench should have decreased. Since the spreading rate to the south would not have been affected ~~significantly~~ to the same extent, then it is expected that the ridge system should have adjusted accordingly. This would then provide the change in the direction of spreading needed for the formation of a leaky transform fault.

However, when the effect of ridge push is included, then the situation gets complicated. Since the elevation of mid-ocean ridges is independent of the spreading rate, then the contribution of ridge push should remain constant regardless of ridge's proximity to a trench. It is quite conceivable that in time, the effect of slab push may eventually dominate over the effect of slab pull. If this is the case, then the rate of spreading and subduction need not significantly change. This is because the slab sinks at some terminal velocity (Forsyth and Uyeda, 1975) and a major change in the stress field is needed before large changes in the rate of sea-floor spreading can be seen.

If in the case of the northeast Pacific, that slab push became the dominant driving mechanism, then a change in the rate of spreading would not occur until the elevation of the ridge was affected. In this case, a change in rate would occur immediately prior to ridge subduction. Therefore, the time at which the ridge would reach the trench would correspond to a change in spreading rate. This change can be seen in analyzing the data north of the Mendocino fracture zone and that south of the Murray fracture zone since these areas are believed to have reached the trench at some later date than the initial contact (ie. the initial contact between the ridge and the trench would affect the regions close by according to the rigid plate hypothesis) (Atwater, 1970).

DATA

An analysis of the rate of spreading along selected areas of the northeast Pacific is now presented. This region is shown in Fig. 1. Magnetic data from Navy map NP-9 was used to determine the spreading rate. In areas not covered by this map, the magnetic lineations identified by Pitman et al. (1974) was used. The Pitman et al. map was used with caution as no comparison with real magnetic data could be accomplished. In areas of overlap between the two maps, some small scale differences were found in the placement of the magnetic anomalies.

In obtaining the ages of these magnetic anomalies, the time scale of LaBreque et al. (1977) was used (Table 2). Previous studies have generally used the time scales of Heirtzler er al. (1968) (Table 3), Pitman et al. (1974) and/or Blakely (1974) (only for the Miocene). The choice of time scale is very important as the spreading rates are directly determined by it. As an example of this, the difference between the LaBreque et al. and the Blakely time scales can be seen in Fig. 25 In obtaining the theoretical magnetic signature of various lines, apparent changes in the rate of spreading are obtained if the time scale is in error. Therefore, small localized deviations are discounted as being not pertinent to this study. We are generally only interested in gross changes in the rate of sea-floor spreading. The data is presented in figures 25 to 45. The synthetic magnetic model was computed using the computer program of W. Jason Morgan of Princeton University.

TABLE 2 REVISED MAGNETIC POLARITY TIME SCALE
FOR CENOZOIC AND LATE CRETACEOUS TIME

Normal polarity intervals (m.y.)	Magnetic anomaly no.	Normal polarity intervals (m.y)	Magnetic anomaly no.
0.00-0.70	1	22.96-23.13	6C
0.89-0.95		23.25-23.49	6C
1.62-1.83	2	23.75-23.92	6C
2.41-2.84	2A	25.24-25.35	7
2.91-3.00	2A	25.42-25.72	7
3.10-3.32	2A	26.14-26.33	7A
3.76-3.85	3	26.63-26.71	8
3.97-4.10	3	26.79-27.54	8
4.24-4.31	3	27.96-28.56	9
4.40-4.59	3	28.62-29.04	9
5.12-5.29	3A	29.57-29.88	10
5.43-5.62	3A	29.94-30.19	10
6.06-6.18		31.11-31.47	11
6.37-6.44	4	31.53-31.96	11
6.51-6.90	4	32.37-32.82	12
6.97-7.02	4	35.26-35.45	13
7.39-7.68	4A	35.52-35.86	13
7.87-7.95	4A	37.26-37.48	15
8.15-8.23		37.50-37.71	15
8.34-8.50	5	38.14-38.38	16
8.54-8.86	5	38.55-38.84	16
8.87-9.15	5	38.89-39.31	16
9.18-9.46	5	39.60-40.52	17
9.48-9.74	5	40.59-40.80	17
9.86-9.91		40.87-41.22	17
10.36-10.43		41.40-41.85	18
10.91-11.09	5A	41.93-42.37	18
11.22-11.49	5A	42.44-42.88	18
11.84-11.87		43.77-44.24	19
11.96-12.01		44.85-46.40	20
12.23-12.41		49.04-50.67	21
12.61-12.87		52.31-53.00	22
13.11-13.52		54.29-54.44	23
13.64-14.11		54.50-55.13	23
14.32-14.42	5B	55.58-55.81	24
14.59-14.73	5B	56.11-56.60	24
15.72-16.02	5C	58.67-59.16	25
16.06-16.24	5C	59.97-60.41	26
16.31-16.50	5C	62.30-62.72	27
17.11-17.45	5D	63.34-64.03	28
17.67-17.69	5D	64.34-64.90	29
18.13-18.67	5E	65.37-66.76	30
18.95-20.07	6	66.84-67.57	31
20.52-20.80	6A	69.20-69.43	32
21.03-21.36	6A	69.65-71.00	32
21.56-21.72		71.34-71.38	
21.92-22.02		71.62-76.48	33
22.24-22.65	6B	79.65-108.19	34

Note: Short wavelength magnetic anomalies that have been dated as events shorter than 40,000 yr have been included; however, their nature still is in question (see text).

From LaBreque et al. (1977)

Figure 24
The geomagnetic time scale. From left to right: Phanerozoic time scale for geologic eras, numbers assigned to bodies and magnetic anomalies, geomagnetic field polarity with normal polarity periods colored black

Table 3
Intervals of normal polarity (m.y.).

0.00– 0.69	11.93–12.43	25.25–25.43	40.03–40.25	67.77–68.51
0.89– 1.93	12.72–13.09	26.86–26.98	40.71–40.97	68.84–69.44
1.78– 1.93	13.29–13.71	27.05–27.37	41.15–41.46	69.93–71.12
2.48– 2.93	13.96–14.28	27.83–28.03	41.52–41.96	71.22–72.11
3.06– 3.37	14.51–14.82	28.35–28.44	42.28–43.26	74.17–74.30
4.04– 4.22	14.98–15.45	28.52–29.33	43.34–43.56	74.64–76.33
4.35– 4.53	15.71–16.00	29.78–30.42	43.64–44.01	
4.66– 4.77	16.03–16.41	30.48–30.93	44.21–44.69	
4.81– 5.01	17.33–17.80	31.50–31.84	44.77–45.24	
5.61– 5.88	17.83–18.02	31.90–32.17	45.32–45.79	
5.96– 6.24	18.91–19.26	33.16–33.55	46.76–47.26	
6.57– 6.70	19.62–19.96	33.61–34.07	47.91–49.58	
6.91– 7.00	20.19–21.31	34.52–35.00	52.41–54.16	
7.07– 7.46	21.65–21.91	37.61–37.82	55.92–56.66	
7.51– 7.55	22.17–22.64	37.89–38.26	58.04–58.94	
7.91– 8.28	22.90–23.08	38.68–38.77	59.43–59.69	
8.37– 8.51	23.29–23.40	38.83–38.92	60.01–60.53	
8.79– 9.94	23.63–24.07	39.03–39.11	62.75–63.28	
10.77–11.14	24.41–24.59	39.42–39.47	64.14–64.62	
11.72–11.85	24.82–24.97	39.77–40.00	66.65–67.10	

Both from Heirtzler et al. (1968)

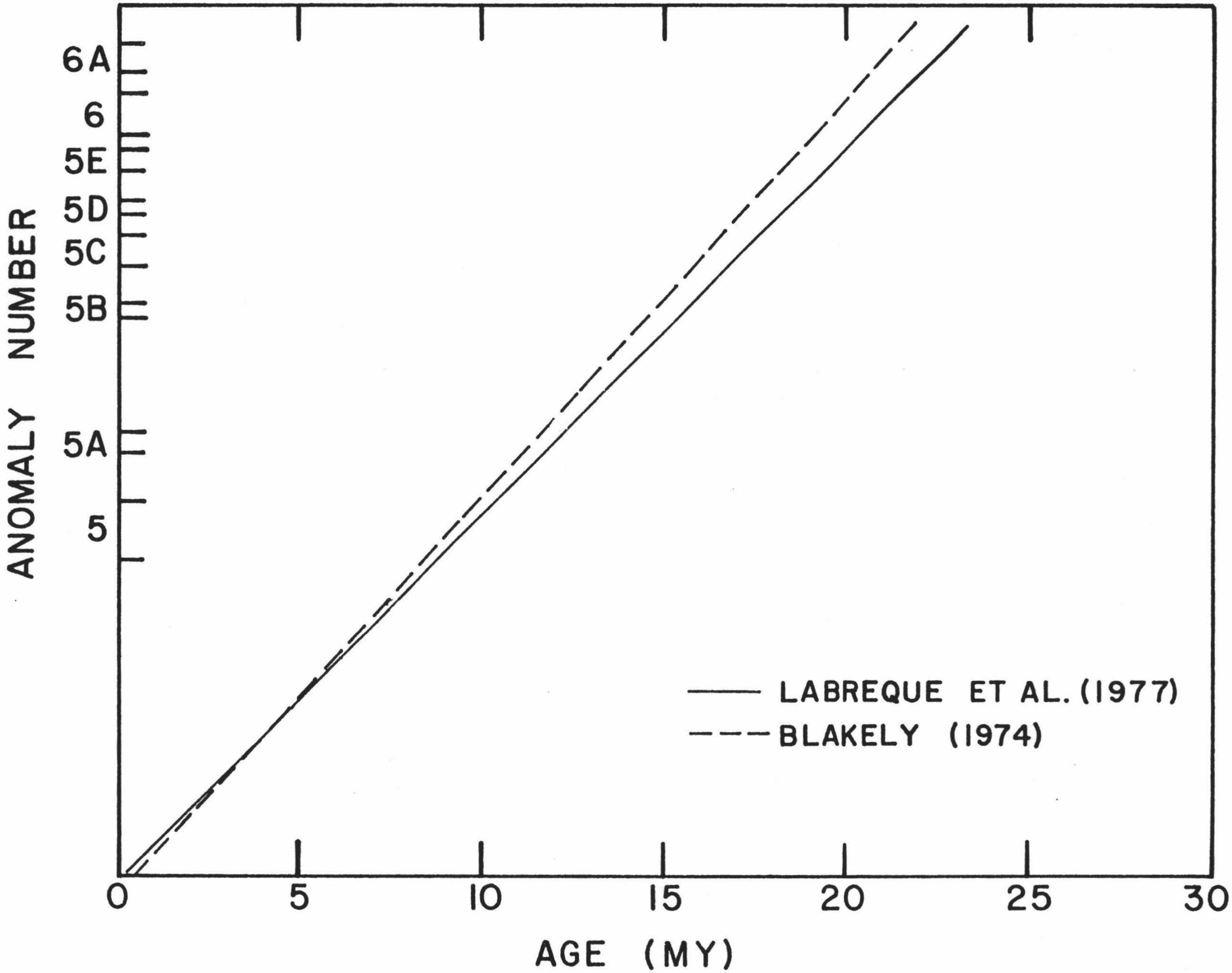


Fig. 25. The LaBreque et al. scale was used as the standard (straight line plot).

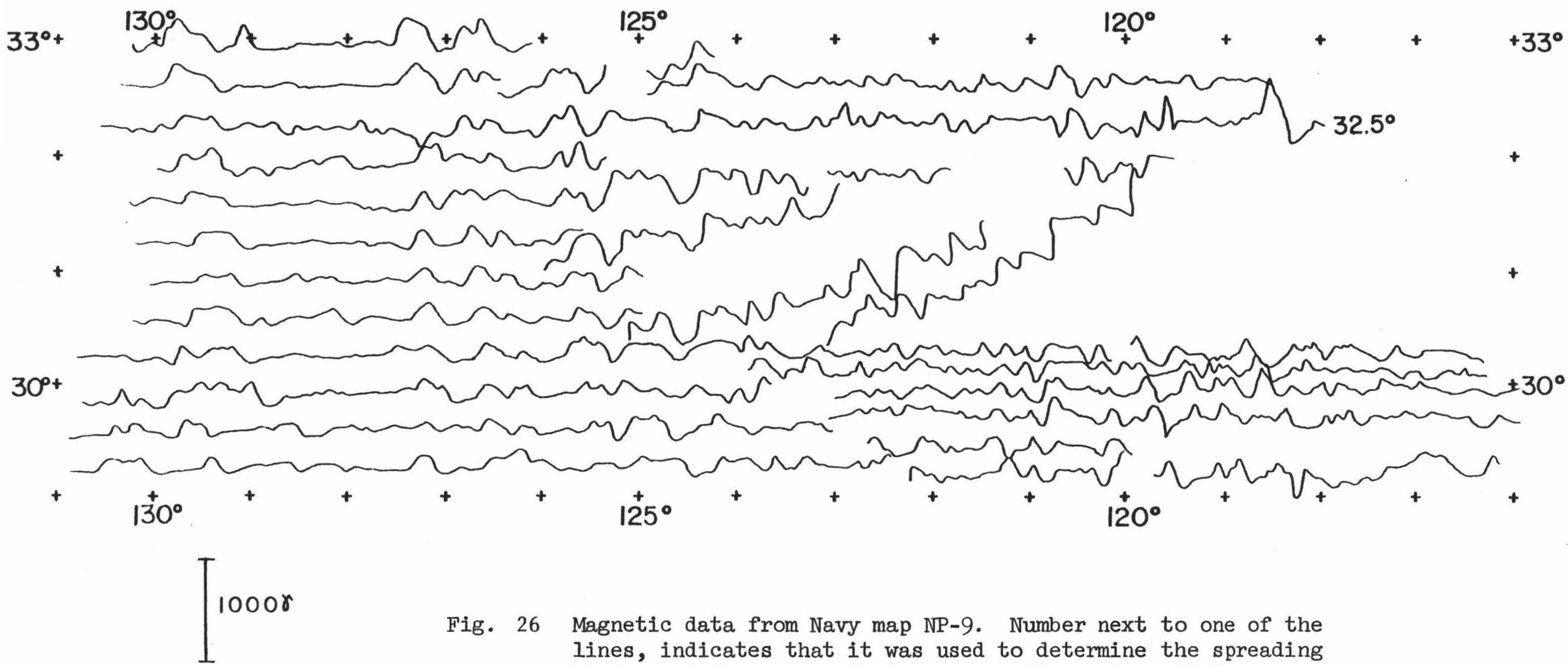


Fig. 26 Magnetic data from Navy map NP-9. Number next to one of the lines, indicates that it was used to determine the spreading history of the area. In this case, it was line 32.5.

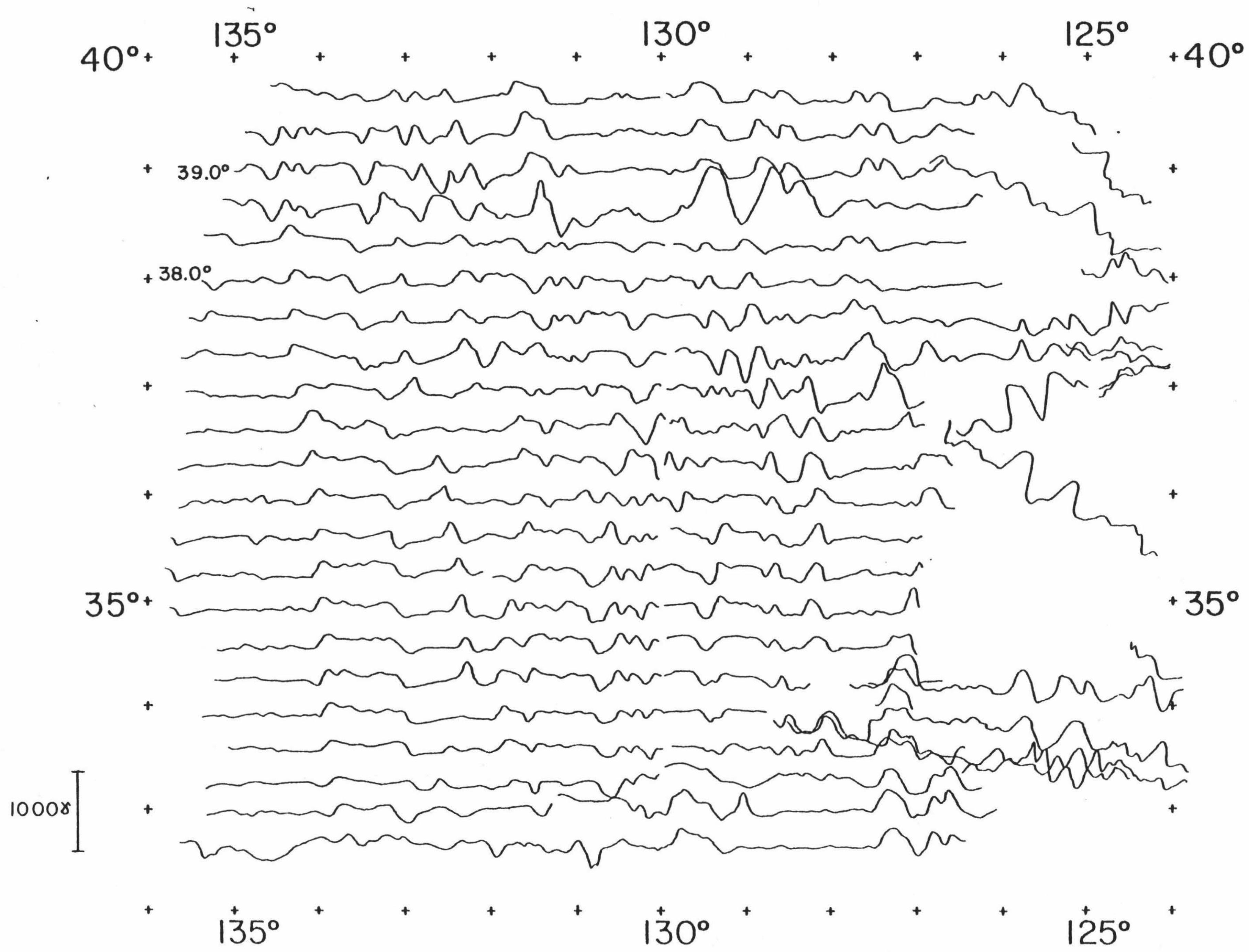


Fig. 27 Magnetic data from Navy map NP-9.

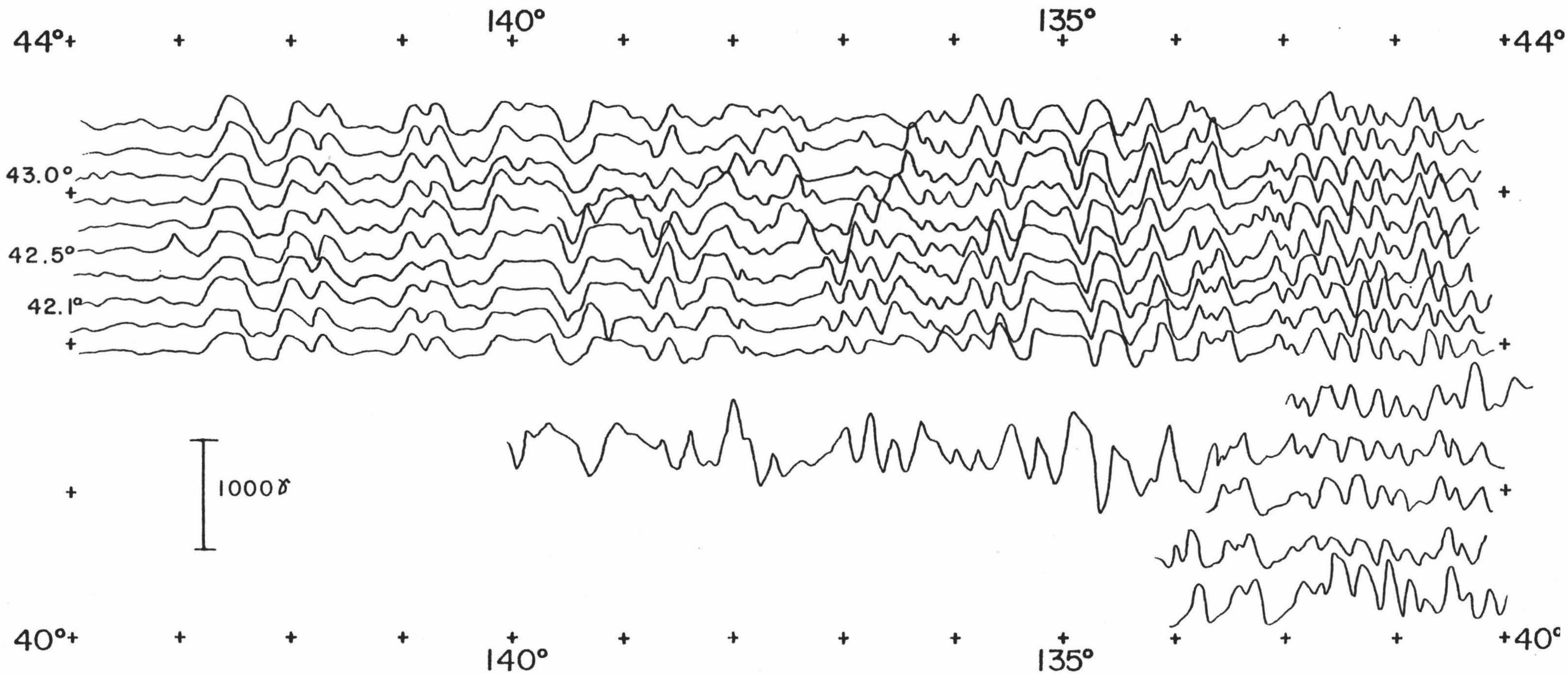


Fig. 28 Magnetic data from Navy map NP-9.

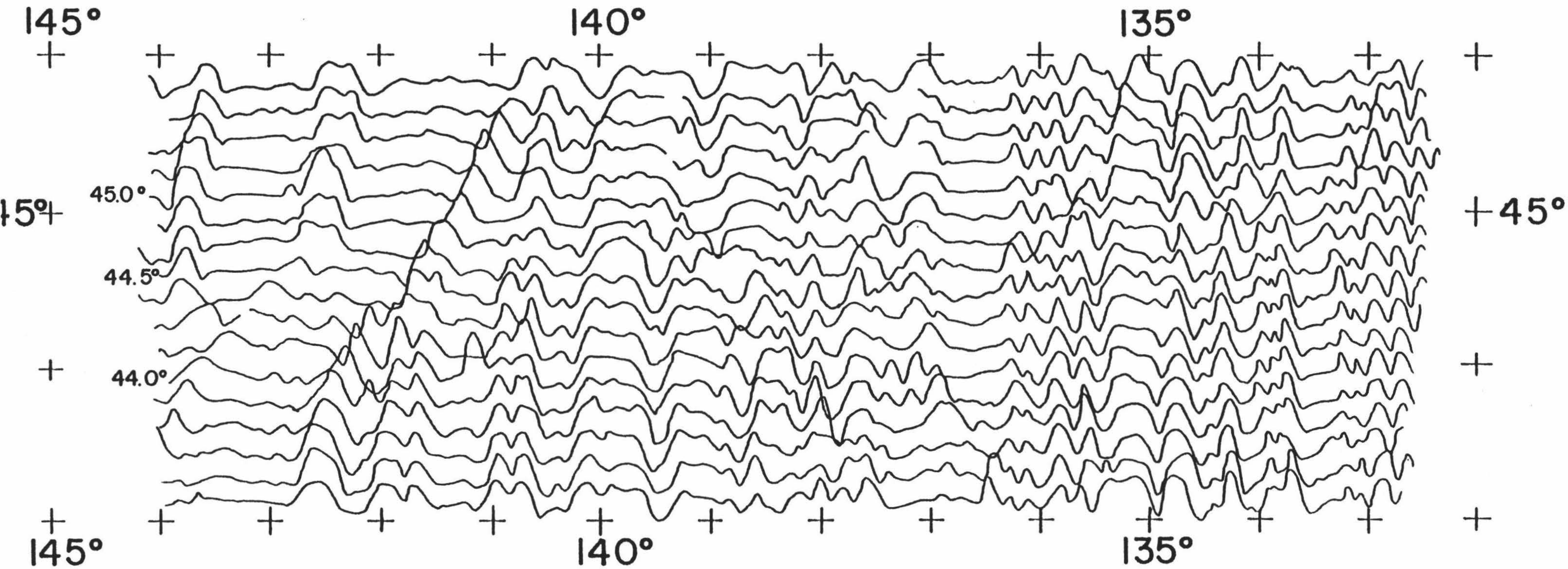
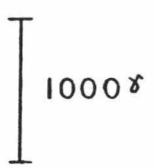


Fig. 29 Magnetic data from Navy map NP-9.



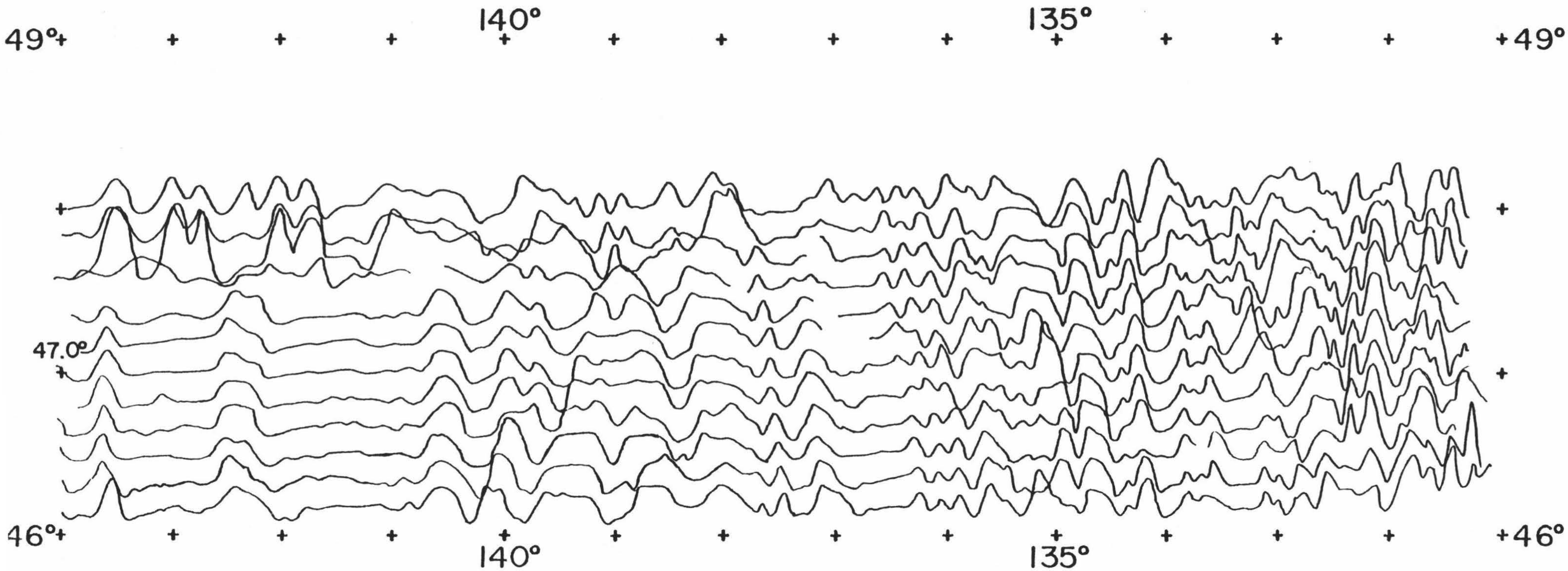
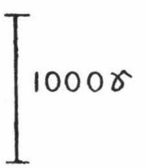


Fig. 30 Magnetic data from Navy map NP-9.



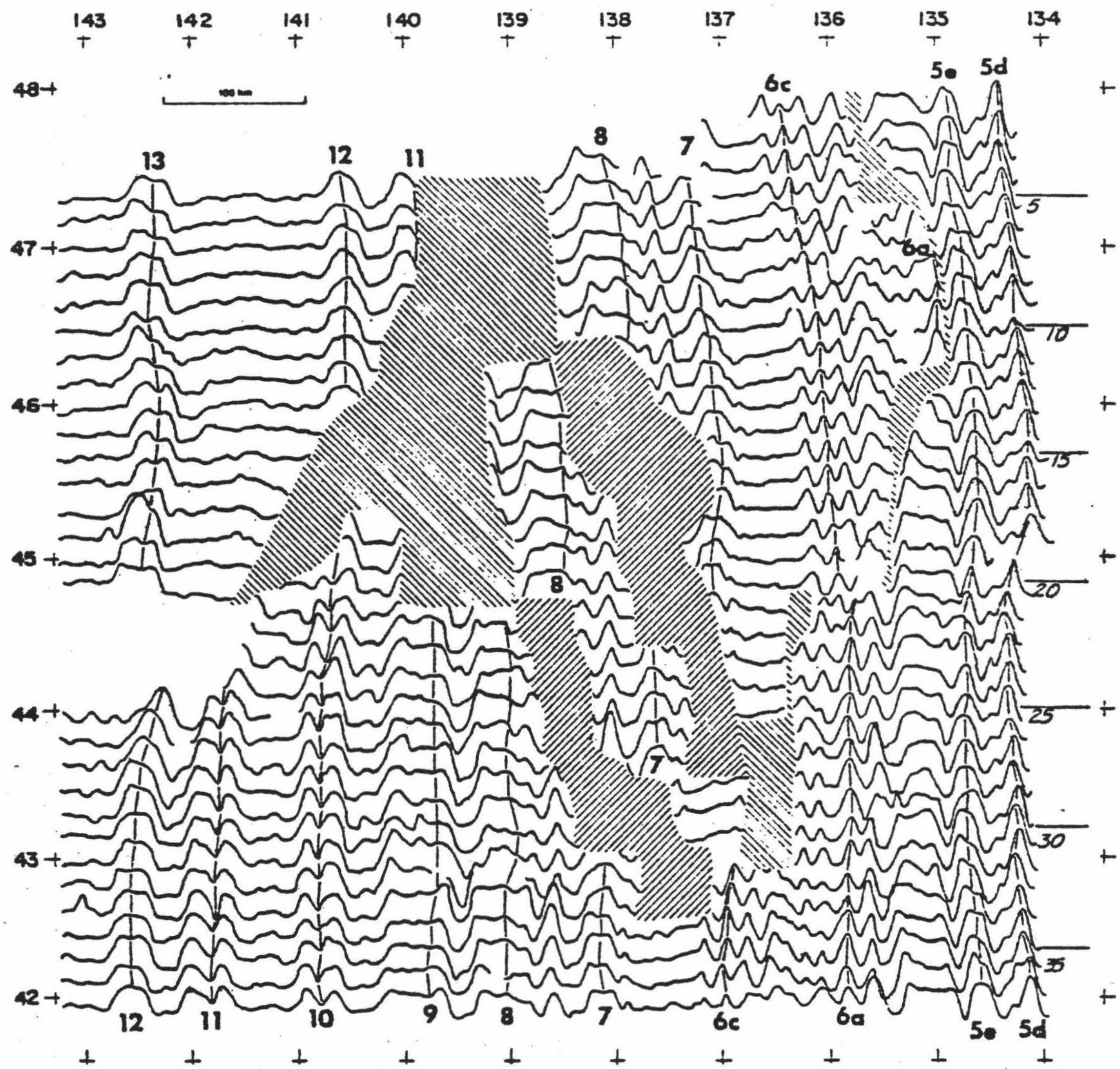


Fig. 31 Magnetic lineations in the Northeast Pacific within the area of study. Anomalies are those identified by Shih and Molnar (1975). Shaded region are areas which have been affected by ridge jumps and/or propagating rifts. Figure is from Shih and Molnar (1975).

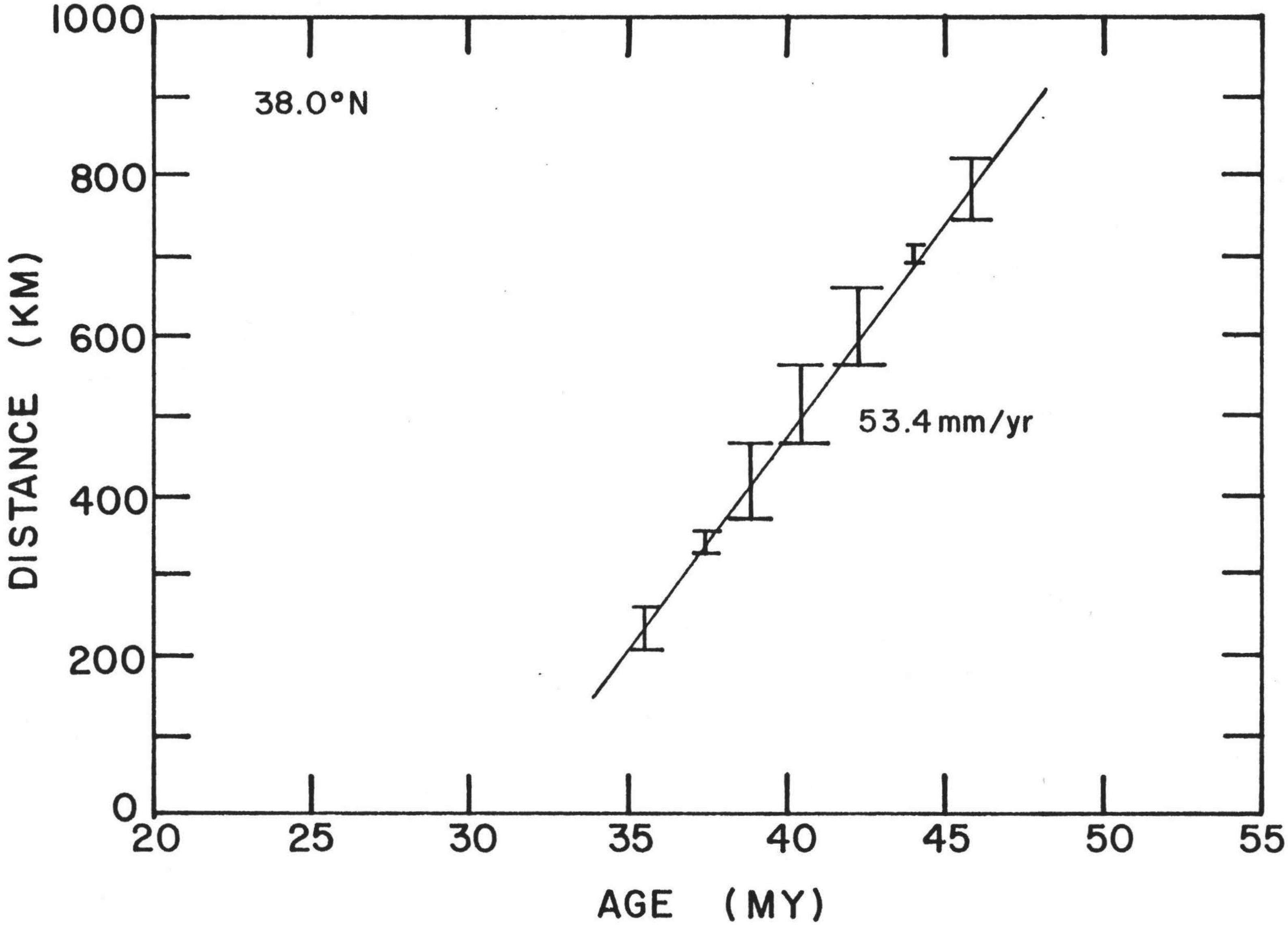


Fig. 32. Distance versus age plot with an estimated spreading rate slope drawn in. See figure 47 for a better analysis of this line.

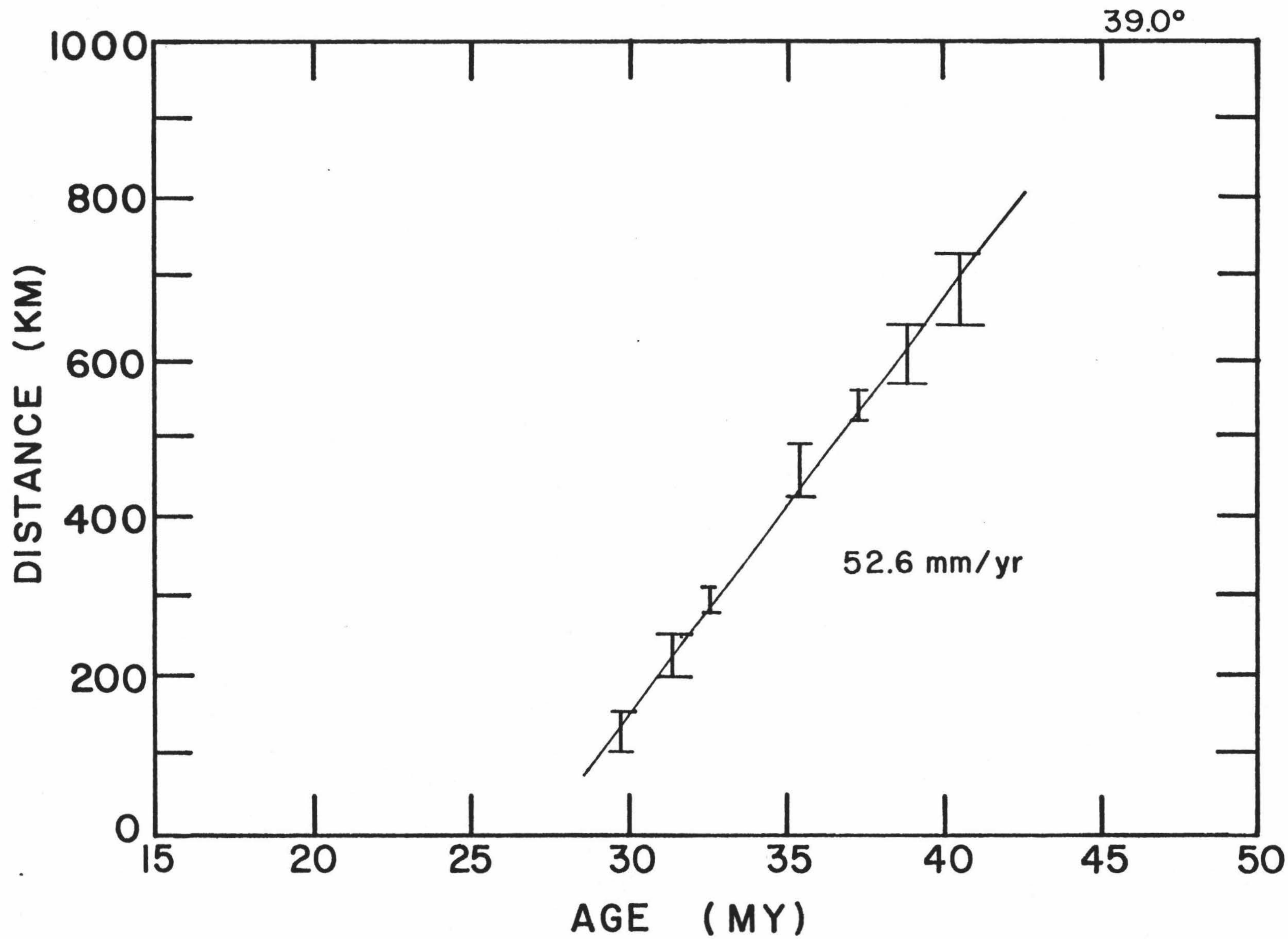


Fig. 33. See figure 44 for a better analysis of this line.

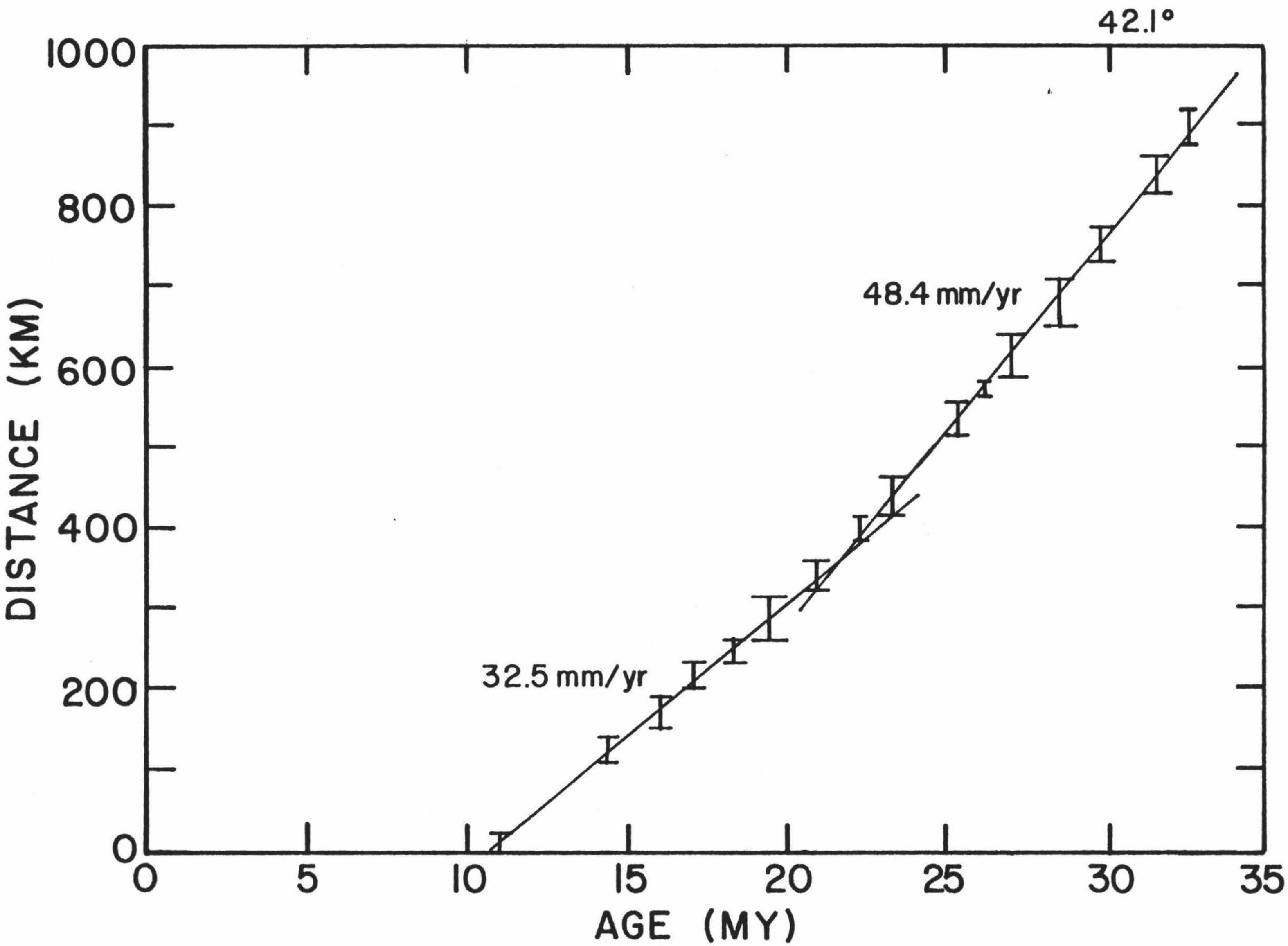


Fig. 34. See figure 45 for a more detailed analysis of this line.

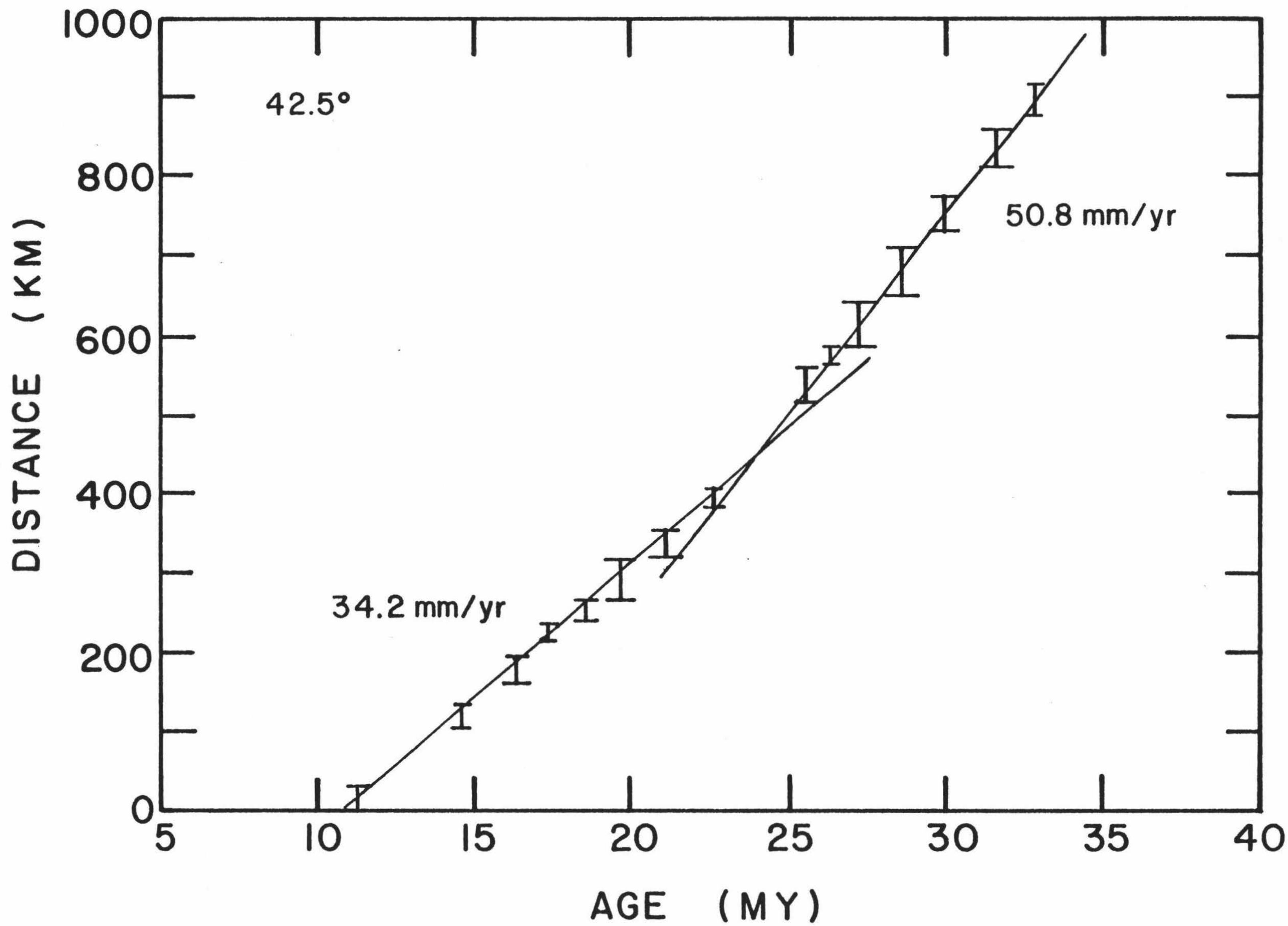
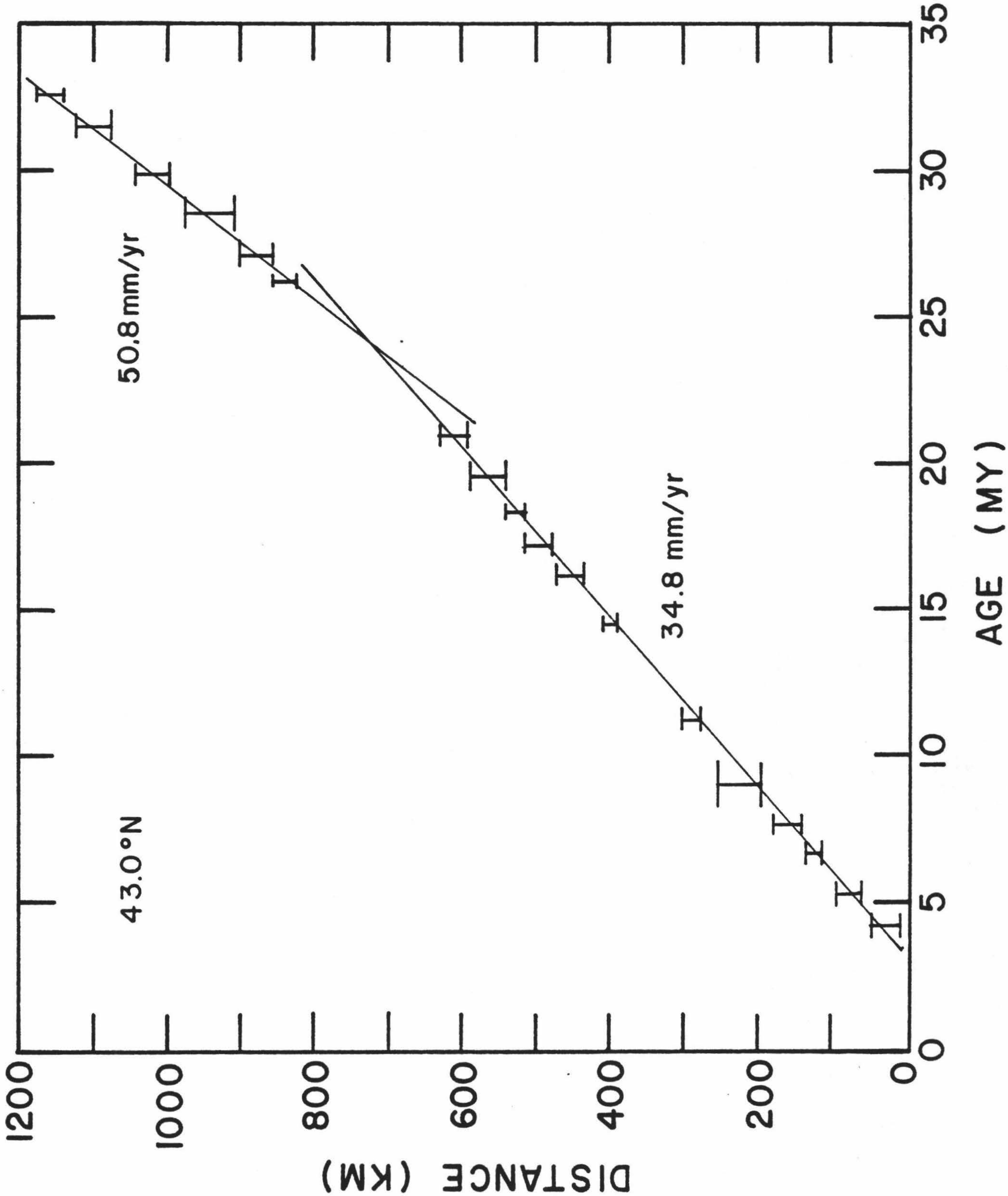

1000
800
600
400
200
0
5
10
15
20
25
30
35
40
DISTANCE (KM)
AGE (MY)
42.5°
50.8 mm/yr
34.2 mm/yr

Figure 35.



43.0°N
50.8 mm/yr
34.8 mm/yr
AGE (MY)
DISTANCE (KM)
0
5
10
15
20
25
30
35
0
200
400
600
800
1000
1200

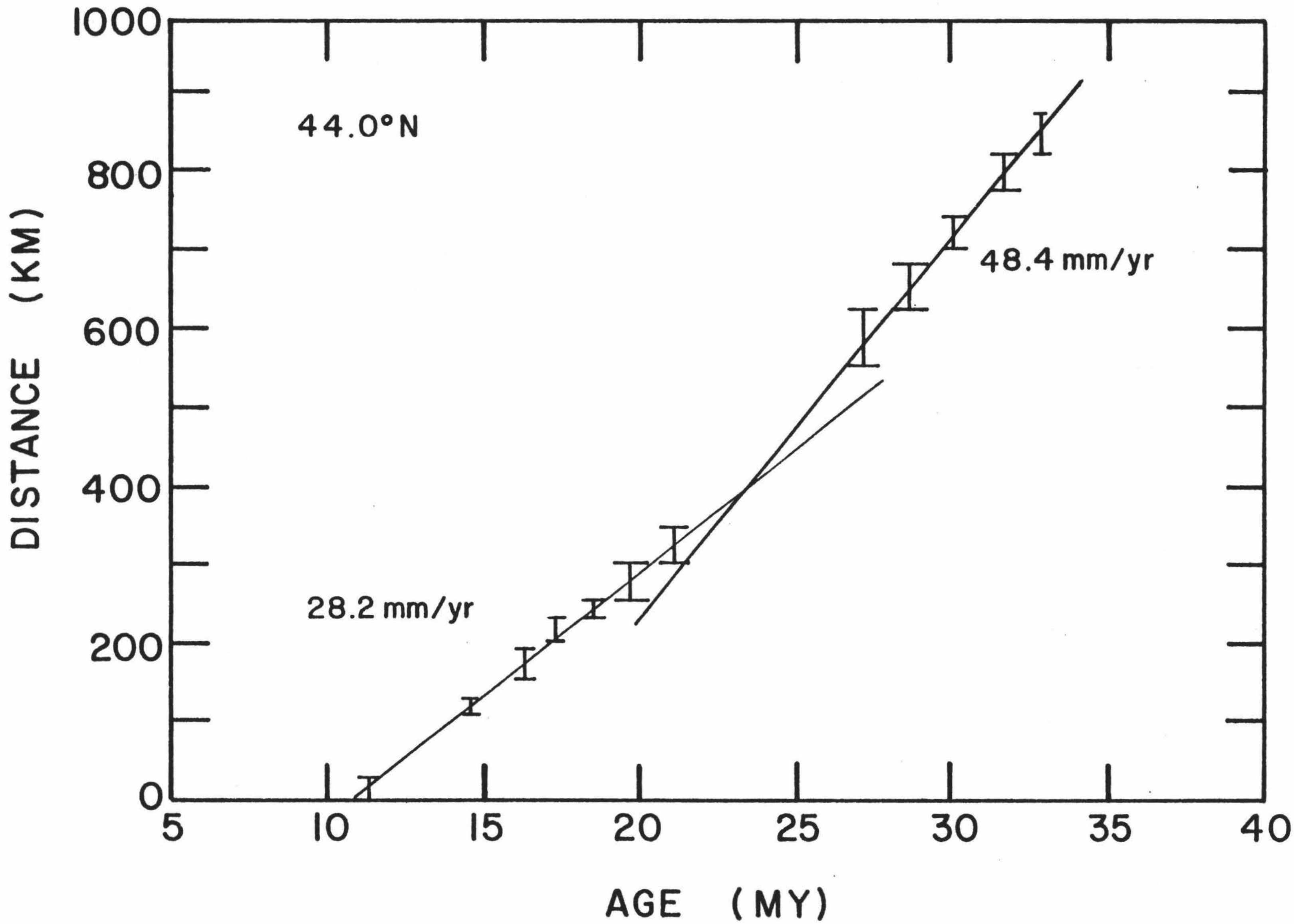

44.0°N
48.4 mm/yr
28.2 mm/yr
DISTANCE (KM)
AGE (MY)
0
200
400
600
800
1000
5
10
15
20
25
30
35
40

Figure 37.

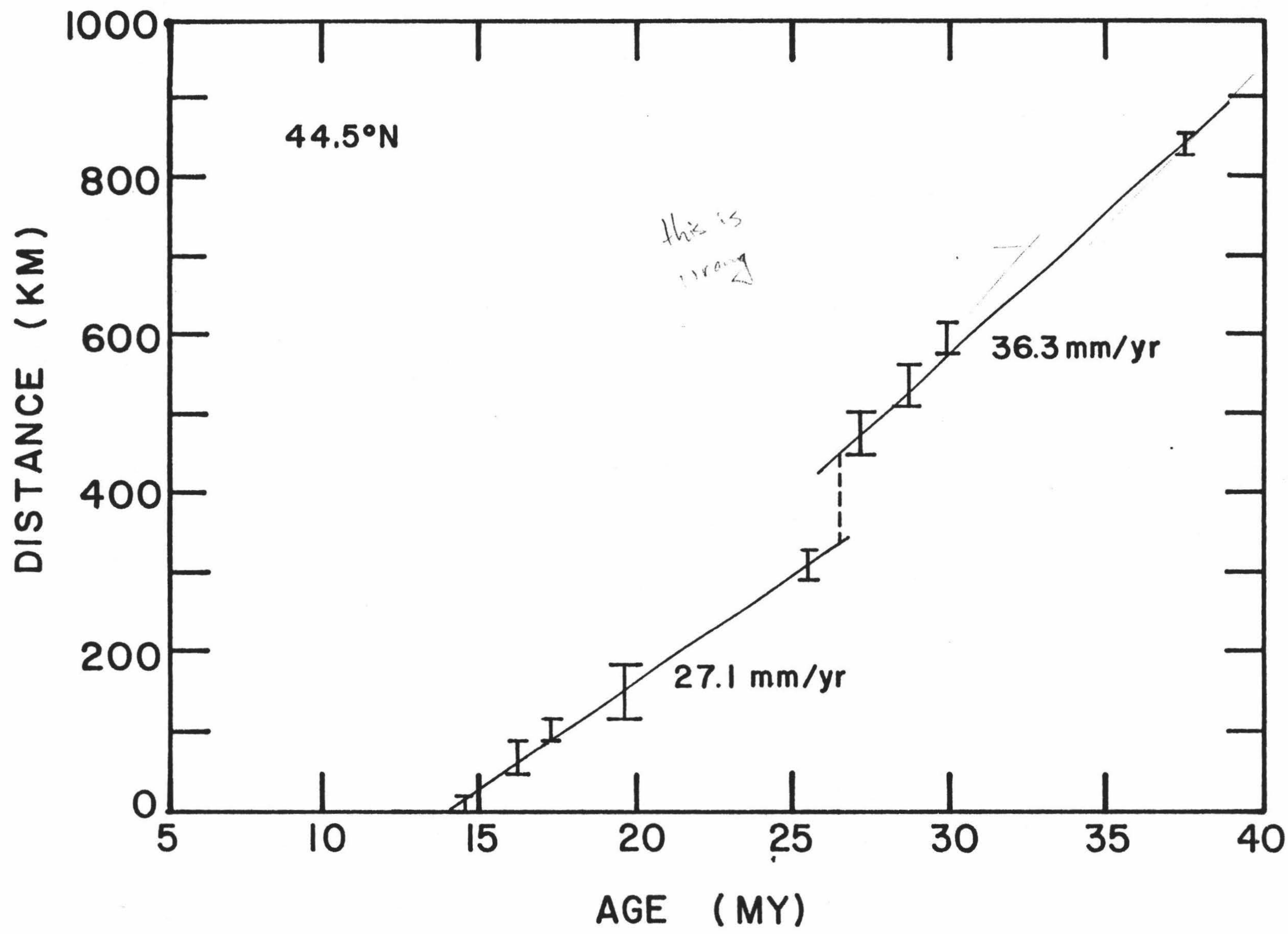


Fig. 38. Dotted line represents a possible ridge jump (onto the Farallon plate).

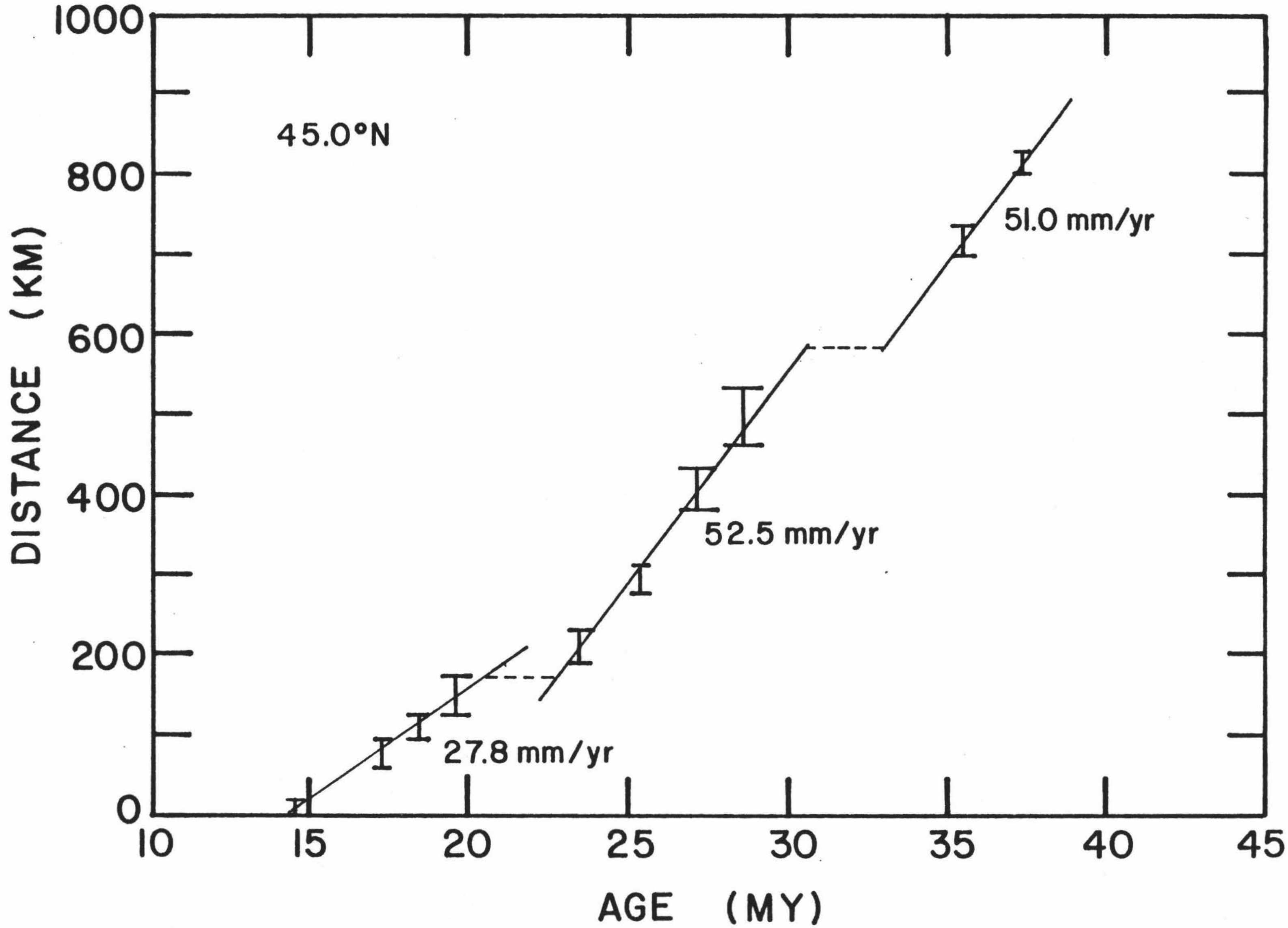


Fig. 39. Dotted lines represent a possible ridge jump onto the Pacific Plate.

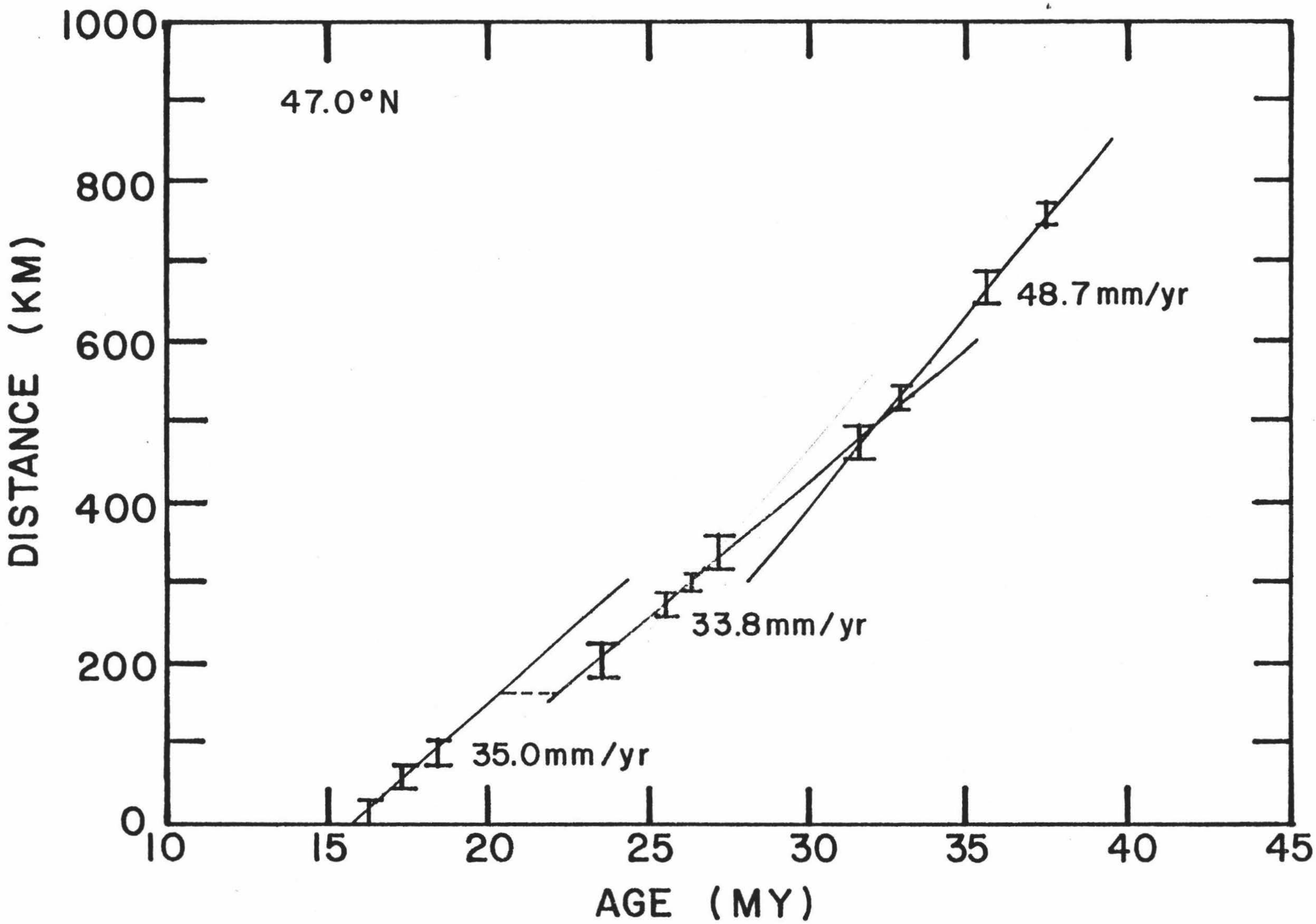
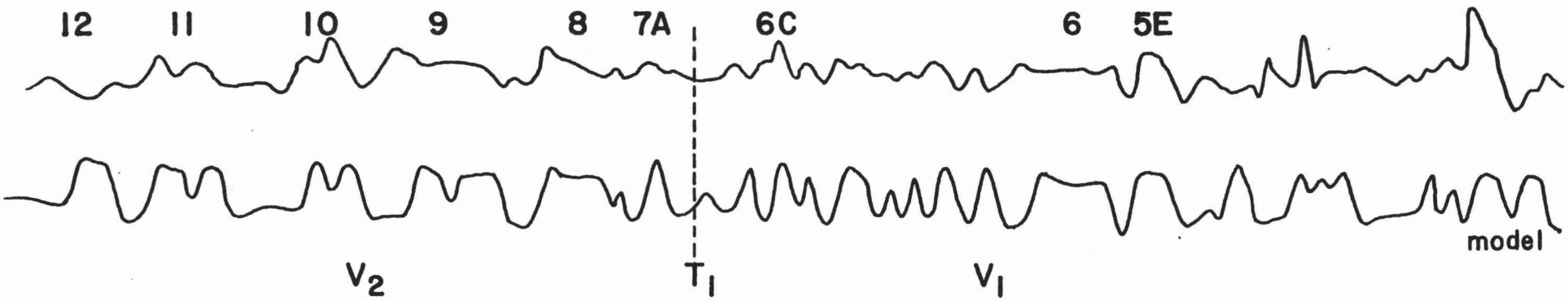


Fig. 40. Dotted lines represent a possible ridge jump onto the Pacific Plate.



REGION BETWEEN
MURRAY AND MOLOKAI
FRACTURE ZONES

LAT. = 32.5° N
LONG. = 123.0° W

1000 γ

100 KM

V_1 = 43.7 mm/yr

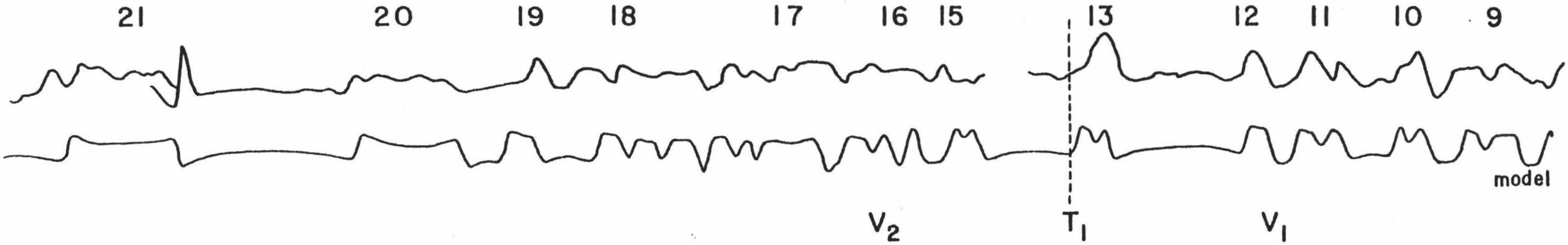
V_2 = 52.1 mm/yr

T_1 = 24.4 mybp

RIDGE JUMP ONTO
PACIFIC PLATE

TIME OF JUMP = 24.4 mybp
DISTANCE = 67 km

Figure 41. T_1 is the time at which the spreading rate changes from V_2 to V_1.



REGION BETWEEN
MURRAY AND PIONEER
FRACTURE ZONES

V_1 = 46.6 mm/yr
V_2 = 56.4 mm/yr
T_1 = 35.9 mybp

LAT. = 34.0°N
LONG. = 129.0° W

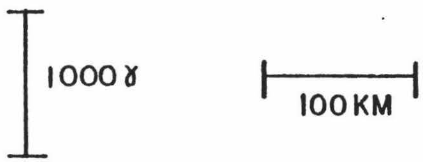


Figure 42.

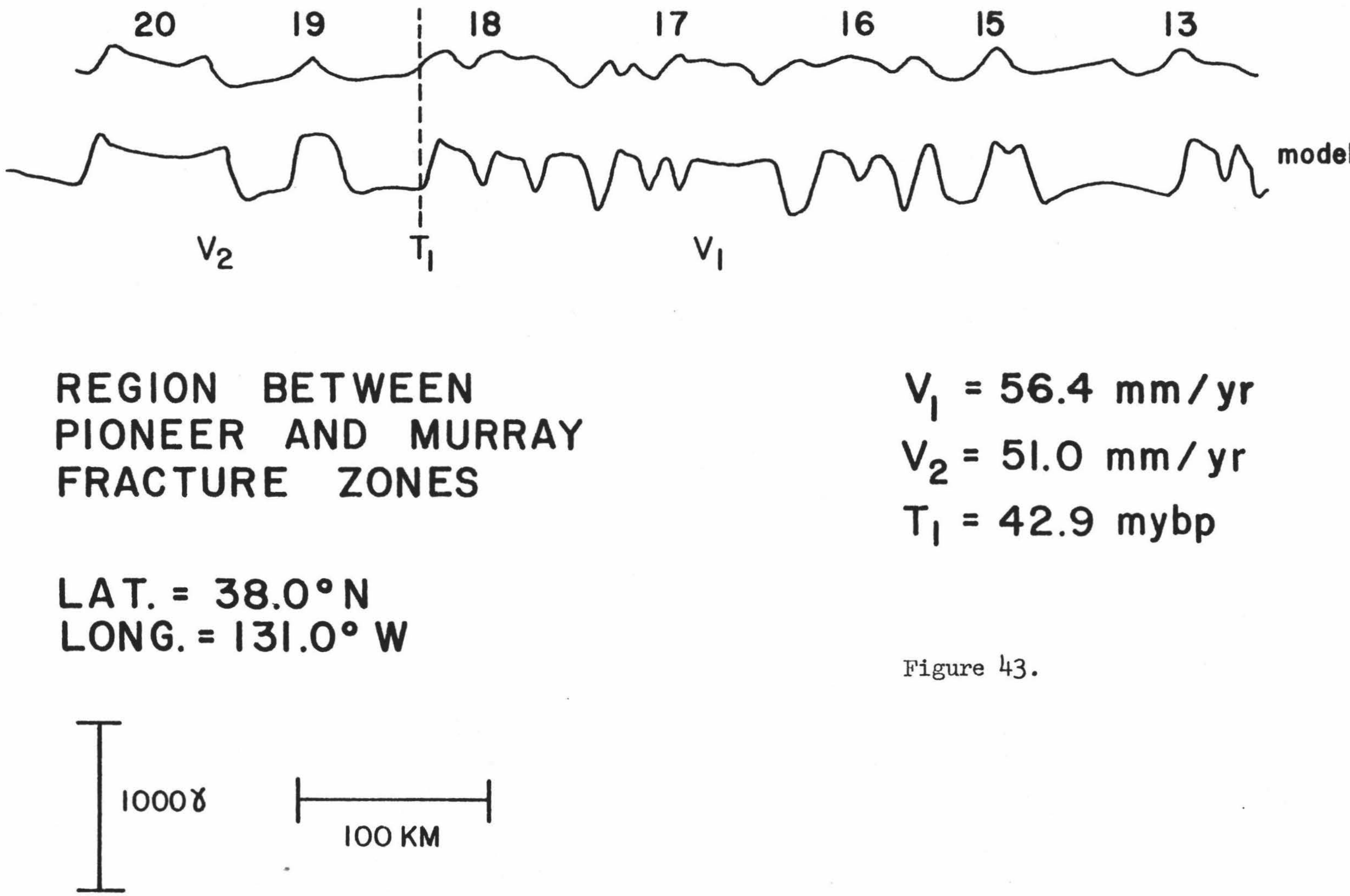
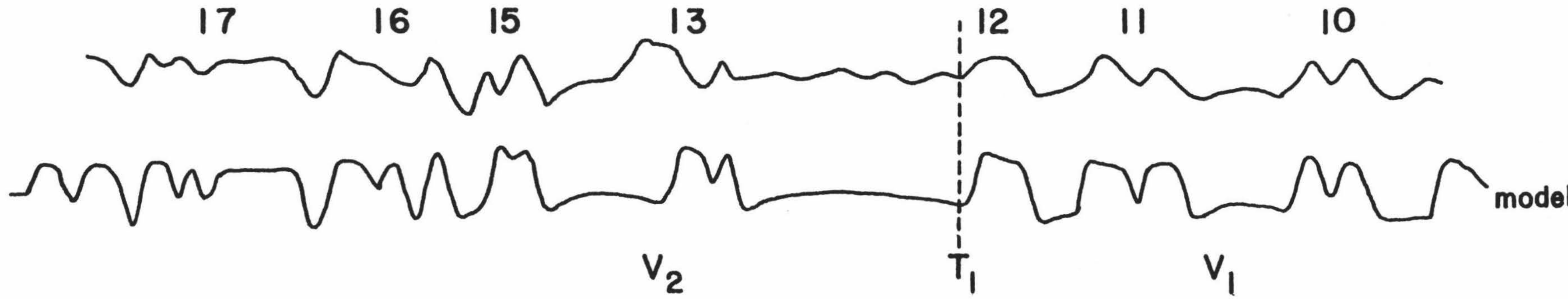

20
19
18
17
16
15
13
model
V_2
T_1
V_1
REGION BETWEEN
PIONEER AND MURRAY
FRACTURE ZONES
LAT. = 38.0° N
LONG. = 131.0° W
V_1 = 56.4 mm/yr
V_2 = 51.0 mm/yr
T_1 = 42.9 mybp
1000γ
100 KM

Figure 43.

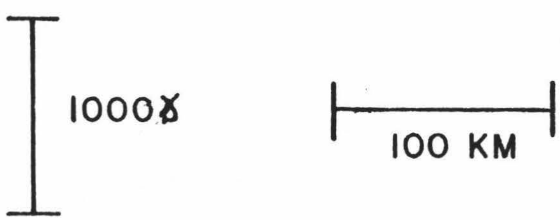


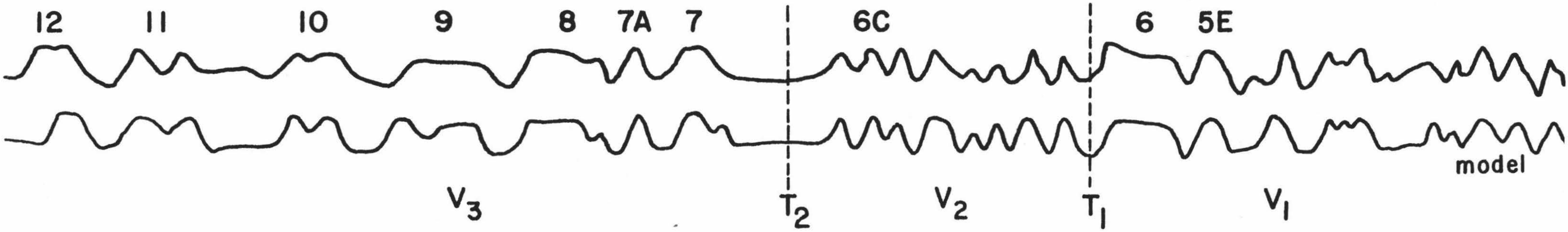
REGION BETWEEN
MENDOCINO AND PIONEER
FRACTURE ZONES

V_1 = 61.6 mm/yr
V_2 = 50.1 mm/yr
T_1 = 32.8 mybp

LAT. = 39.0° N
LONG. = 131.0° W

Figure 44.





REGION BETWEEN
MENDOCINO AND SURVEYOR
FRACTURE ZONES

LAT. = 42.1° N
LONG. = 137° W

V_1 = 34.3 mm/yr
V_2 = 40.1 mm/yr
V_3 = 50.4 mm/yr
T_1 = 20.1 mybp
T_2 = 24.5 mybp

1000 γ

100 KM

Figure 45.

DISCUSSION

Several problems are readily observed form the magnetic data. The spreading history for the region between the Murray and the Pioneer fracture zones is not simple and may reflect upon the amount of intraplate deformation that may have occurred (Fig. 46). For this reson, data from this region will be used with caution. However, the predicted intraplate deformation apparently took place. Due to age differential across the Mendocino fracture zone, the area to the north remained relatively stable while the region to the immediate south fractured up. If the break-up occurred at the fracture zones, then several individual small plates are thereby formed. These plates in turn can thus behave semi-independently of each other. This explains for the lack of correlation in the magnetic data.

If the plate between the Murray and the Mendocino fracture zones did break up, then determining the spreading history of this area is nearly impossible. It is too difficult to take into account any combination of possible ridge jumps, ridge migrations and/or asymmetrical spreading. It is known that the area to the west of the EPR would remain relatively stable while the region to the east would not. But the extent of this is unknown.

I therefore choose to accept the spreading history derived from north of the Mendocino fracture zone simply because it would have been less affected by intraplate deformation as the ridge approached the trench. The disturbed region shows the effects of the change in the stress field but certain sections appear to have remained relatively intact. If this is the case, then the only major spreading rate change occurred at around 24 myBP. This matches remarkably well with a selected line between the Murray and the Molokai fracture zones (see line at 32.5°N).

An attempt can now be made to explain the tectonic history of the region. Since the Cretaceous period, two mid-ocean ridges, the Kula-Farallon and the Kula-Pacific were subducted, thereby leaving only the EPR. As the EPR approached the North American trench (rate of subduction faster than the rate of spreading), the rheological properties of the oceanic lithosphere should have changed. This is clearly shown through the amount of intraplate deformation that occurred.

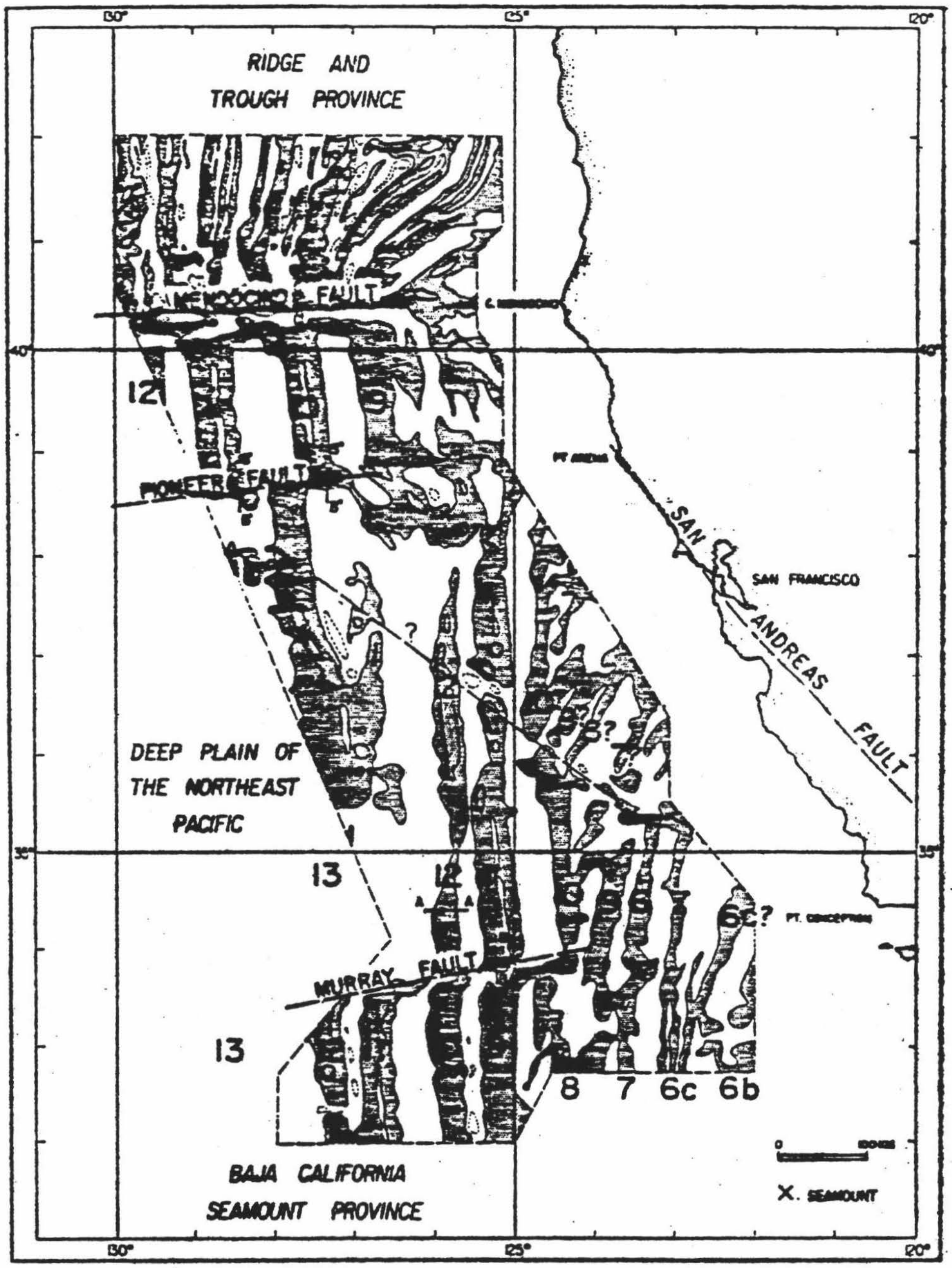


Fig. 46. Magnetic lineations off the coast of California are shown here. Note the fractured nature in the region between the Murray and the Pioneer Faults. Anomalies are numbered using the system of Heirtzler et al. (1968). Figure is from McKenzie and Morgan (1969).

The model proposed by Atwater (1970) had the EPR reaching the NA trench at around 31 myBP. This date is based upon the Heirtzler et al. (1968) time scale and can therefore be updated to 28 myBP if the newer LaBreque et al. (1977) time scale is used instead. Nonetheless, the spreding history does not show any significant change at this time. It is possible that some time shift between the cause and the effect, either forward or backwards existed thus explaining the lack of correlation.

More significant is the observation that the predicted decrease in the spreading rate prior to ridge subduction did not occur. This would tend to indicate that either 1) slab pull is not the dominant driving force, 2) some other driving forces were in control, 3) some form of time shift took place, and/or 4) some of the resistive forces systematically decreased.

Exactly why no change in the spreading rate while plate deformation did occur is an enigma. It is very likely that all the plate interactions are highly related and therefore changes in the spreading rate would seriously disrupt the equilibrium that exists. Intraplate deformation on the other hand would not have as much of an effect on a worldwide basis. However, this would imply that when a mid-ocean ridge is subducted that "all hell should break loose." Correlation of orogenic events from ocean to ocean should be readily apparent but is in fact not. Even apparent matches must be viewed with caution since the possibility of a random pairing of events is highly likely (ie. with the amount of tectonic events occurring, some matches even though unrelated must happen). A simple explanation for all this is to allow the continental masses to absorb most of the changes. Since these plates represent resistive features, then this point is not beyond credibility.

To say that slab pull is not the primary driving mechanism based upon the data just presented is highly speculative. The simple fact that it can explain this observation does not obliterate the evidence presented in other studies on plate motion. Instead it may prove to be more advantageous to analyze some other possibility.

Instead of the later age of 28 myBP used by Atwater, this preferred model has the ridge being subducted at 24 myBP. This is the date that corresponds to the change in spreading rate north of the Mendocino fracture zone and the ridge jump and change in rate between the Molokai and the Murray fracture zones. Furthermore, in order to explain the absence of the newer anomalies that would have been formed, the trench would had to have remained for at least four million years longer. Therefore, instead of the formation of RTF and FFT triple junctions (F=transform fault, T=trench, and R=mid-ocean ridge), a RTT and a FTT triple junctions must have existed (Fig. 47). In terms of a simple triple junction analysis (McKenzie and Morgan, 1969), this configuration is stable.

The extra four million years may have been the time for the ridge segment between the Mendocino and Murray fracture zones to have been subducted. This corresponds nicely with another change in the spreading rate observed in the magnetic data (see line at 42.1°N). At this time the trench is no longer needed and a transition to a transform fault is observed. This is shown graphically in figure 48. This may also explain why the San Andreas fault is inland instead of on the oceanic plate. By the time the trench was destroyed, the exinct but still anomalously hot, ridge segment is underneath the North American plate. If the Proto-Gulf of California as proposed by Karig and Jensky (1972) was present, then it would have represented a zone of weakness which was then broken through by the subducting ridge. This model is also compatible with those which have the EPR underlying the North American plate which thereby form the Rocky Mountains (J.T. Wilson, 1978 personal communication).

It is quite conceivable that a mid-ocean ridge could be subducted down into a trench without destroying the trench. Such an occurance has been hypothesized to have occurred in both the Japan-Kuril trench and the Aleutian trench when the Kula-Pacific ridge was destroyed (DeLong et al., 1978; Uyeda and Miyashiro, 1974). This may be due to the subsequent destruction of the thermal balance of the ridge system. If sufficient heat is conducted away (fracturing of the lithosphere with an increase in the amount of hydrothermal circulation of sea water), then the magma chamber under the subducting ridge could be destroyed.

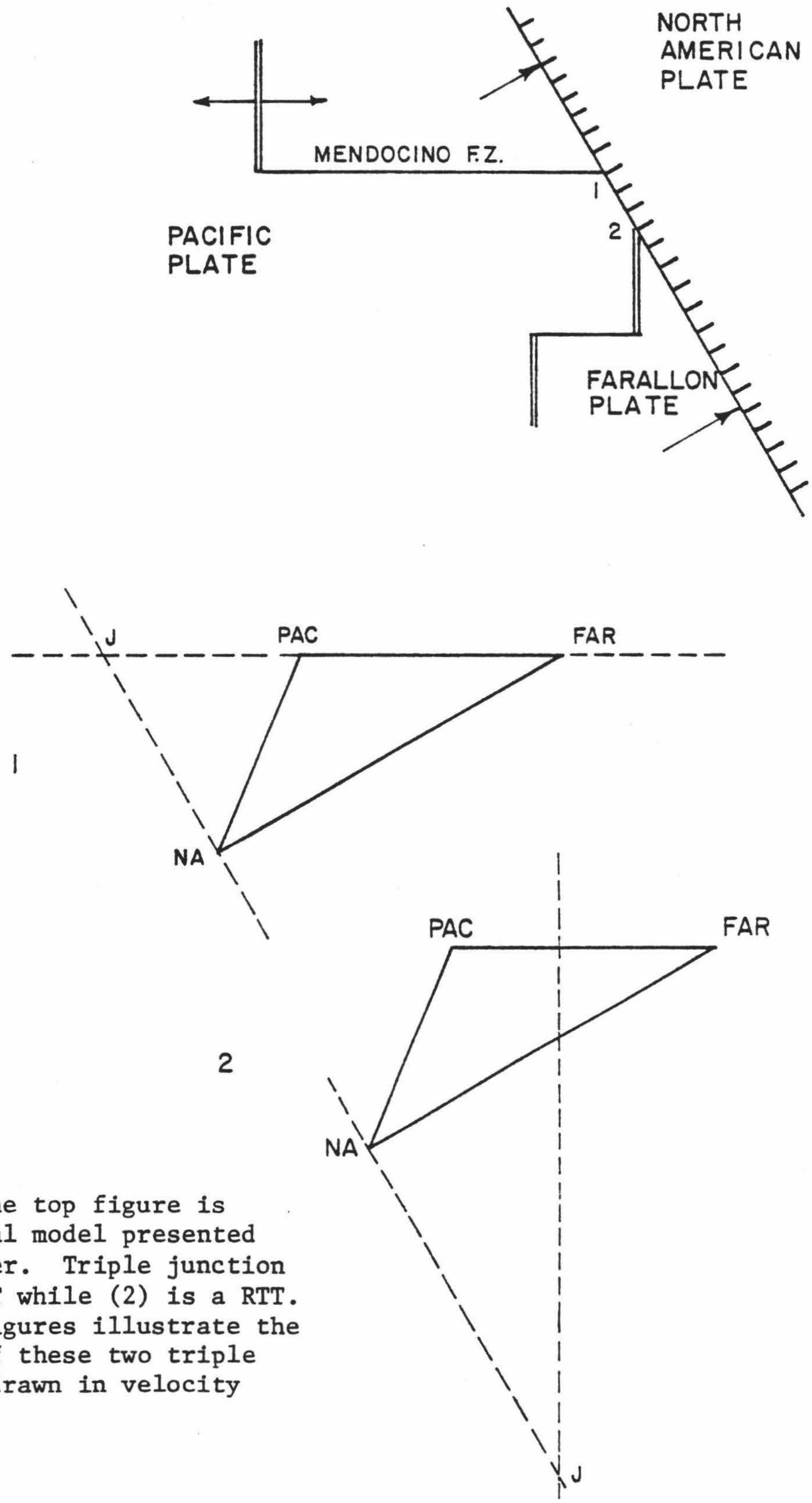


Fig. 47. The top figure is the graphical model presented in this paper. Triple junction (1) is a FTT while (2) is a RTT. Lower two figures illustrate the stability of these two triple junctions (drawn in velocity space).

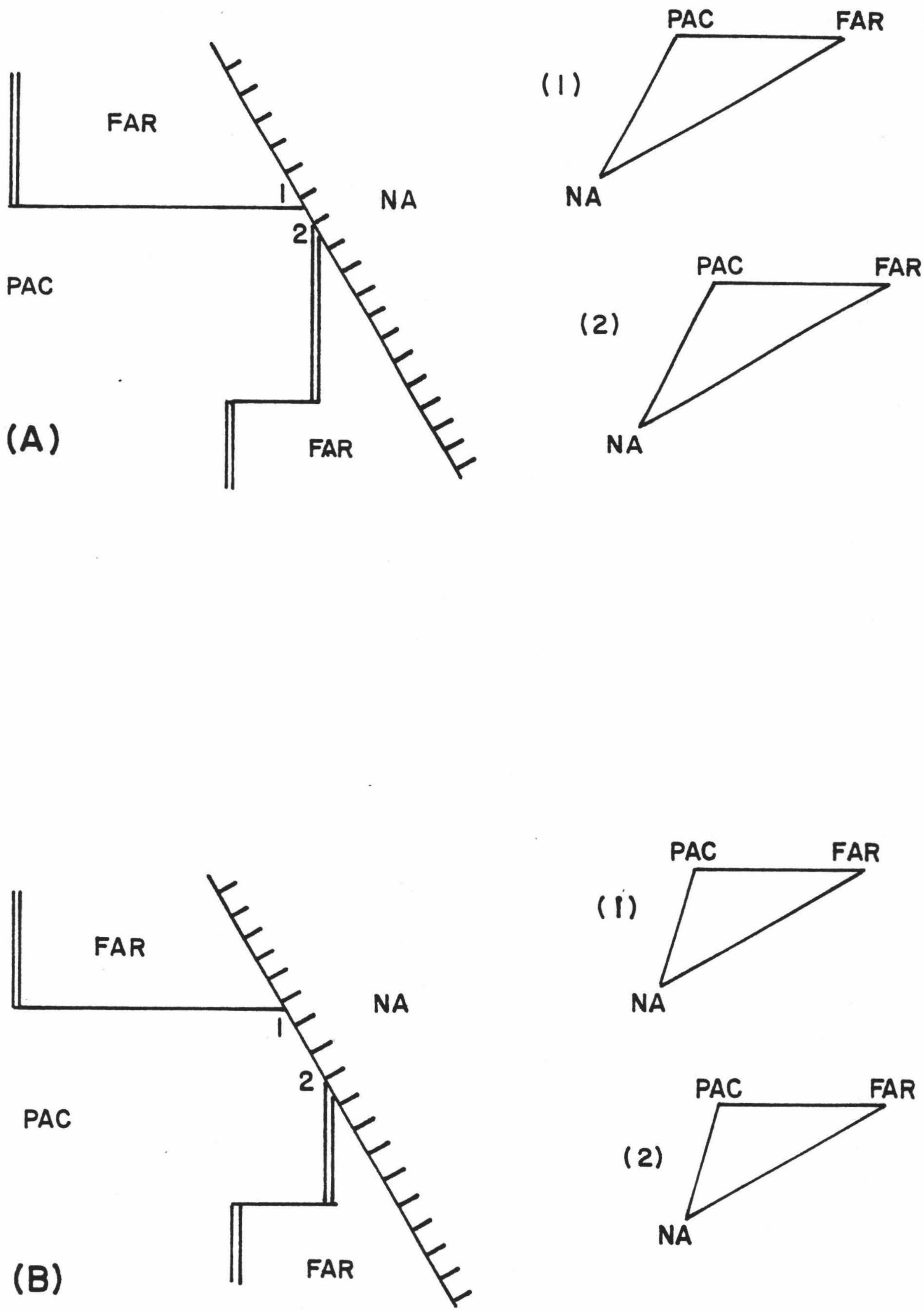


Fig. 48. This figure shows the proposed model for the subduction of the EPR in the region between the Mendocino and the Murray Fracture Zones. Figures on the right are drawn in velocity space. Note the change in the direction of motion between the Pacific and the North American Plates.

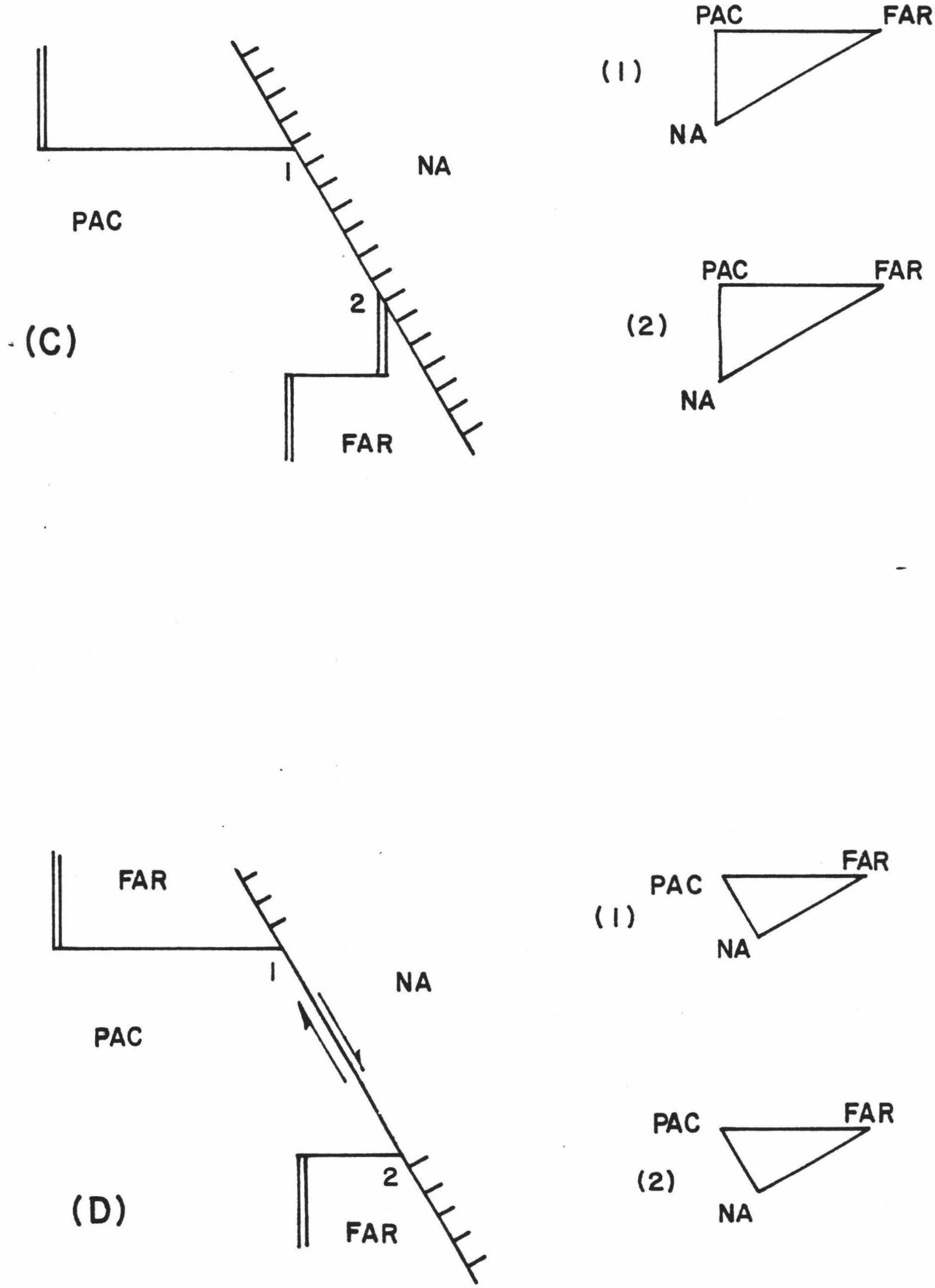


Fig. 48 (continued). Note the transition from a trench to a transform fault (Pacific-North American boundary) as shown by the figures on the right.

CONCLUSION

The tectonic model for the history of the northeast Pacific proposed by Atwater needs to be reevaluated in more detail. In light of the data presented in this paper, the following conclusions and inferences can be made.

The anomalies off the coast of California ~~are~~ may be around 4 m.y. older than the actual time at which the ridge reached the trench. If this is so, then approximately 200 km. of the Pacific plate has been subducted into the former North American trench (4 m.y.*50 km/my). If the dip angle of the downgoing plate was about 45°, then this would put the subducted ridge crest around 140 km. inland by the time the San Andreas fault was formed (200 km.*cos 45°). Furthermore, due to the new time at which the San Andreas fault is proposed to have formed, the tectonic history of California needs to be reevaluated. This can be done through a detailed analysis of the San Andreas fault and secondly, by looking at the volcanic history of the area.

In terms of the structure of mid-ocean ridge systems, it was found that some of the predicted deformation did occur. However, reasons (physical mechanisms) for the deformation process were not determined. It may prove valuable to compare the tectonic history of this region with that of the Aleutians and Japanese Islands. Furthermore, an analysis of the region near the Chile Rise where it is supposedly being subducted needs to be carried out.

Finally, it should be mentioned that there has been no physical model developed for the subduction of a mid-ocean ridge. The most likely method to do such a study would be to set up a thermal model of both the ridge and the trench and combine the two together. Furthermore, a continuation of the leaky transform fault model is still waiting to be developed.

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