SUBLITHOSPHERIC UPWELLING DISTRIBUTION AND ITS
IMPLICATIONS REGARDING HOT SPOTS
AND MANTLE CONVECTION

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ABSTRACT

The spacing of sublithospheric upwellings can be estimated from the interaction of plate velocity, plate thickness, and hot spot distribution. For the major plates, the average area per hot spot exponentially increases with the product of plate velocity (V) and plate thickness (t). The minimum area per hot spot or maximum hot spot density occurs for thin, slowly moving plates. The sublithospheric upwelling density, estimated at Vt=0, is 3.0×10^{6} km^2/upwelling. This spacing implies a depth of convection of about 720 km for three dimensional cells. This depth of upper mantle convection implies that two separate convection systems are present, an upper and a lower. Upper mantle viscosity is constrained by the Rayleigh number and depth of convection to be 1.2×10^{22} poises.

PREFACE

I am grateful to Michael Carr, Jason Morgan, and Claude Herzberg for guidance and criticisms, and to Eric Christofferson, who stirred my interest in hot spots and mantle convection.

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INTRODUCTION

A sublithospheric upwelling can be considered as an area of rising mantle material. These upwellings, either deep mantle plumes (Morgan, 1971; 1972) or shallow convection cells (Richter and Parsons, 1975), create hot spots on the surface of the earth only where they have melted a conduit to the surface. Two factors can prevent rising diapirs from a upwelling from reaching the surface. These are plate thickness and plate velocity. If this simple model is correct and if the sublithospheric upwelling distribution is random, then hot spot density should be related in some way to plate velocity and plate thickness.

A correlation between outbreaks of African volcanism and pauses in the motion of the African plate was suggested by Briden and Gass (1974). Mid-plate volcanism was explained in terms of plate velocity and plate thickness by Gass et al. (1978) and later by Pollack (in press) and Sahagian (1980).

DATA

Hot Spot Distribution

The number of hot spots on each plate is based on the observations and criteria of Burke and Wilson (1976) (table 1 and fig. 1). Hot spots at or very near plate boundaries were divided between the two adjacent plates, since it is unclear to which plate they belong. These hot spots are circled in fig. 1. Although other hot spot lists exist (Frazier, 1979), this one was chosen since it is the most comprehensive.

Plate Velocities

The velocities of the plates (fig. 2) are as calculated by Minster and Jordan (1978; in press) and agree with other studies (Solomon et al., 1977; 1974). The motion of a plate over a mantle upwelling limits the time available for heat transfer through the lithosphere. Lithospheric thinning occurs in moving plates by convective heat transfer (Withjack, 1979; Detrick and Crough, 1978), but only a fraction of the sublithospheric upwellings are able to penetrate to the surface.

Plate Thicknesses

Plate thicknesses (fig. 3) are estimated from heat flow data of sclater et al. (1980). The thicknesses in table 1 are averaged over the area of each plate. These average values are in agreement with those calculated from other models (Kanamori and Press, 1970; Dziewonski, 1971; Okal, 1977; Yoshii, 1975; Chapman and Pollack, 1977; Crough, 1975). The global average lithospheric thickness is estimated at 120 km. Lithospheric thickness

PLATE	HOT SPOTS	PLATE AREA $\times 10^{12}$ m^2	VELOCITY $\times 3.2\times 10^{-10}$ m/sec	THICKNESS km	Vt	A/S $\times 10^{12}$m^2
Eurasia	17.5	68.0	.5	156	78	3.8
Australia	7.5	61.0	6.0	115	690	8.1
Pacific	11.5	113.0	9.0	100	900	9.8
N. America	12	58.8	2.5	136	340	4.9
S. America	4	46.2	3.1	144	446	11.6
Nazca	3	16.4	6.0	65	390	5.5
Africa	35.5	78.4	0-1.5	144	0-216	2.2
Antarctica	17	59.9	.6	133	80	3.5

Table 1

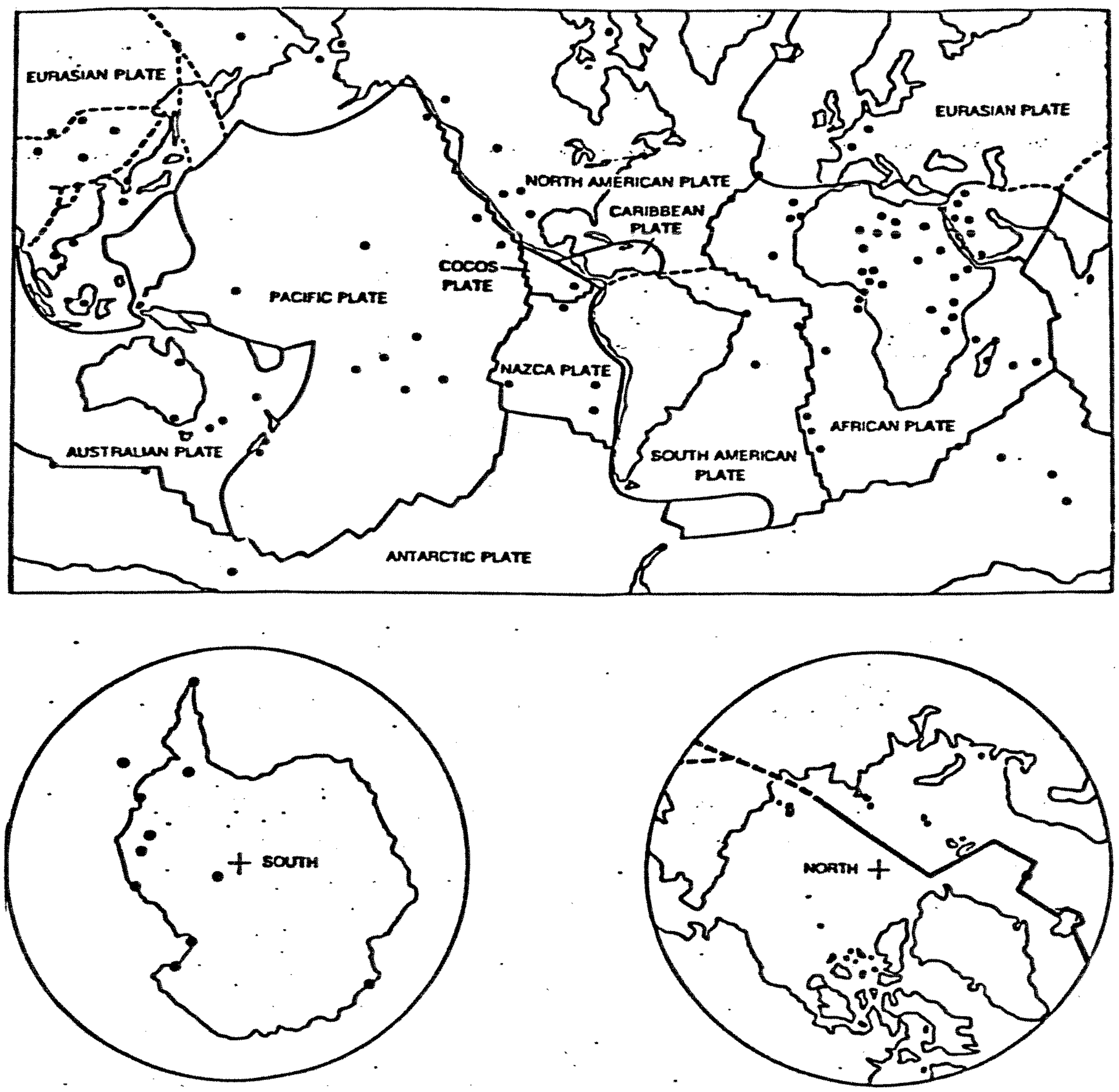


Figure 1

Hot spots on the Earth's surface. From Burke and Wilson, (1976).

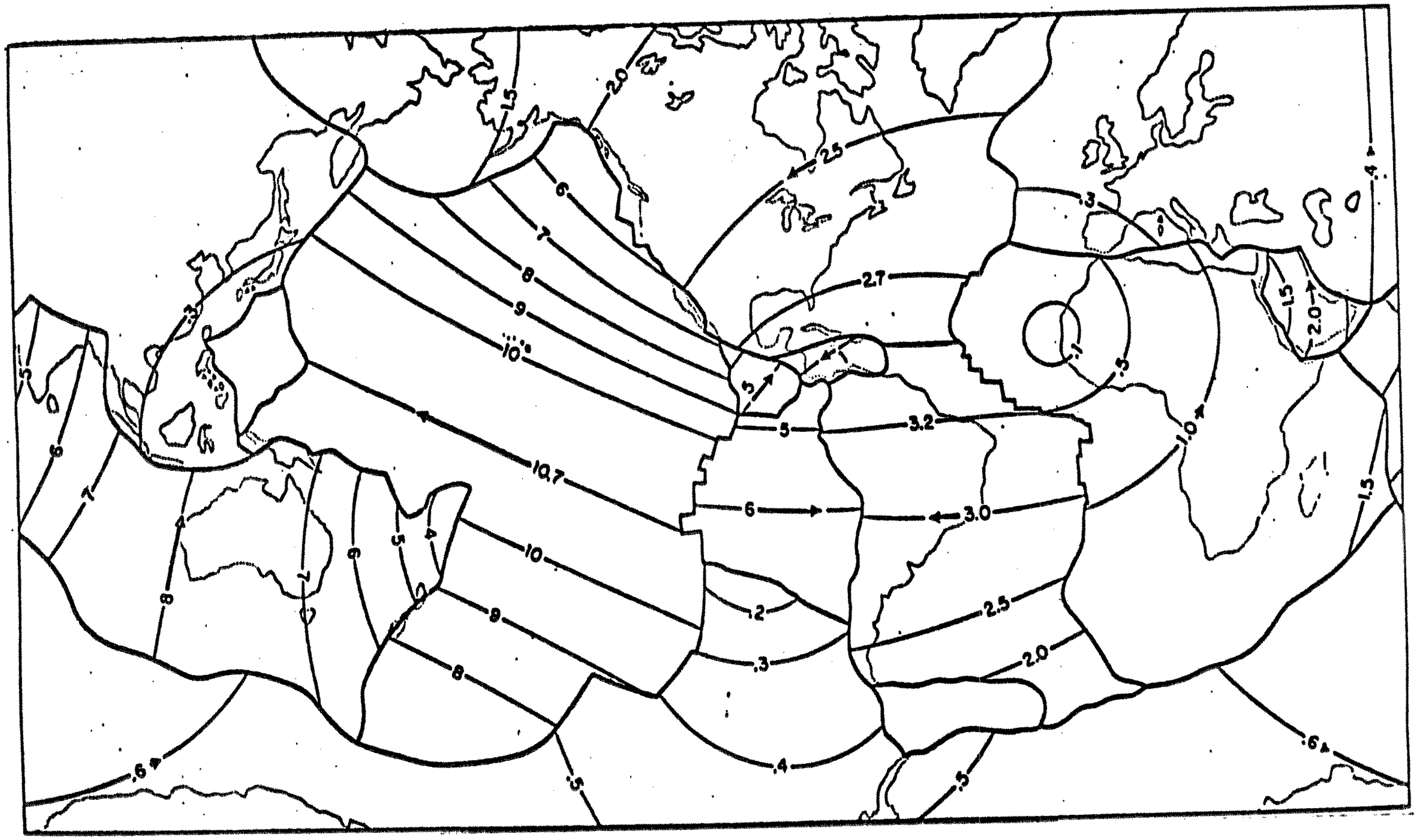


Figure 2

Absloute plate velocities calculated from Minster et al. (in press).

Figure 3

Lithospheric thickness. From Gass et al. (in press).

in part determines the amount of time necessary for an upwelling to penetrate the lithosphere. The thickness depends primarily on the history of the plate involved. For oceanic regions, thickness is directly related to the square root of age (Sclater et al., 1980; Crough and Thompson, 1976).

Rate of Ascent of the Lithosphere-Aesthenosphere Boundary

The rate of ascent of the lithosphere-aesthenosphere boundary as a sublithospheric upwelling impinges on the lithosphere was estimated at $2x10^{-8}$ cm/sec by Withjack (1979). This corresponds to basaltic magma and is assumed constant throughout the earth. The rate of ascent of the lithosphere-aesthenosphere boundary in part determines the amount of time necessary for an upwelling to penetrate the lithosphere.

FORMULATION

The penetration ability of upwellings (the observed hot spot density) depends exponentially upon velocity and thickness because, as a magma loses heat, its temperature falls exponentially with respect to time (Jaeger, 1968) from its initial temperature to the ambient temperature of its surroundings. If the amount of time it takes for a magma to cool and solidify is longer than the time necessary for the magma to penetrate the lithosphere, there will be volcanism at the surface. Since upwellings under any plate are not likely to be all of the same strength (radius, temperature) but are randomly distributed about some mean, some upwellings may fail to reach the surface even if the average upwelling would easily penetrate the plate. Similarly, some exceptionally strong upwellings will be able to penetrate even thick, fast moving plates.

The sublithospheric upwelling density, K (expressed as area per upwelling) can be inferred using the following relation:

$$A/S = K \exp(cVT/R) \qquad (1)$$

where A is plate area, S is the number of hot spots on the plate, V is plate velocity with respect to a fixed mesospheric frame, T=t/g is the thickness of the plate divided by the global average lithospheric thickness, R is the rate of ascent of the lithosphere-aesthenosphere boundary, and c=.10 is a constant of proportionality (Sahagian, 1980).

An exponential relationship is indicated by the plot of ln(A/S) vs. Vt (fig. 4). This dependence can be further verified by examining the limiting cases of velocity, thickness, and R. As $V \to 0$, or $t \to 0$ or $R \to \infty$ the lithosphere is effectively stripped away, and what remains is the sublithospheric area per upwelling, K.

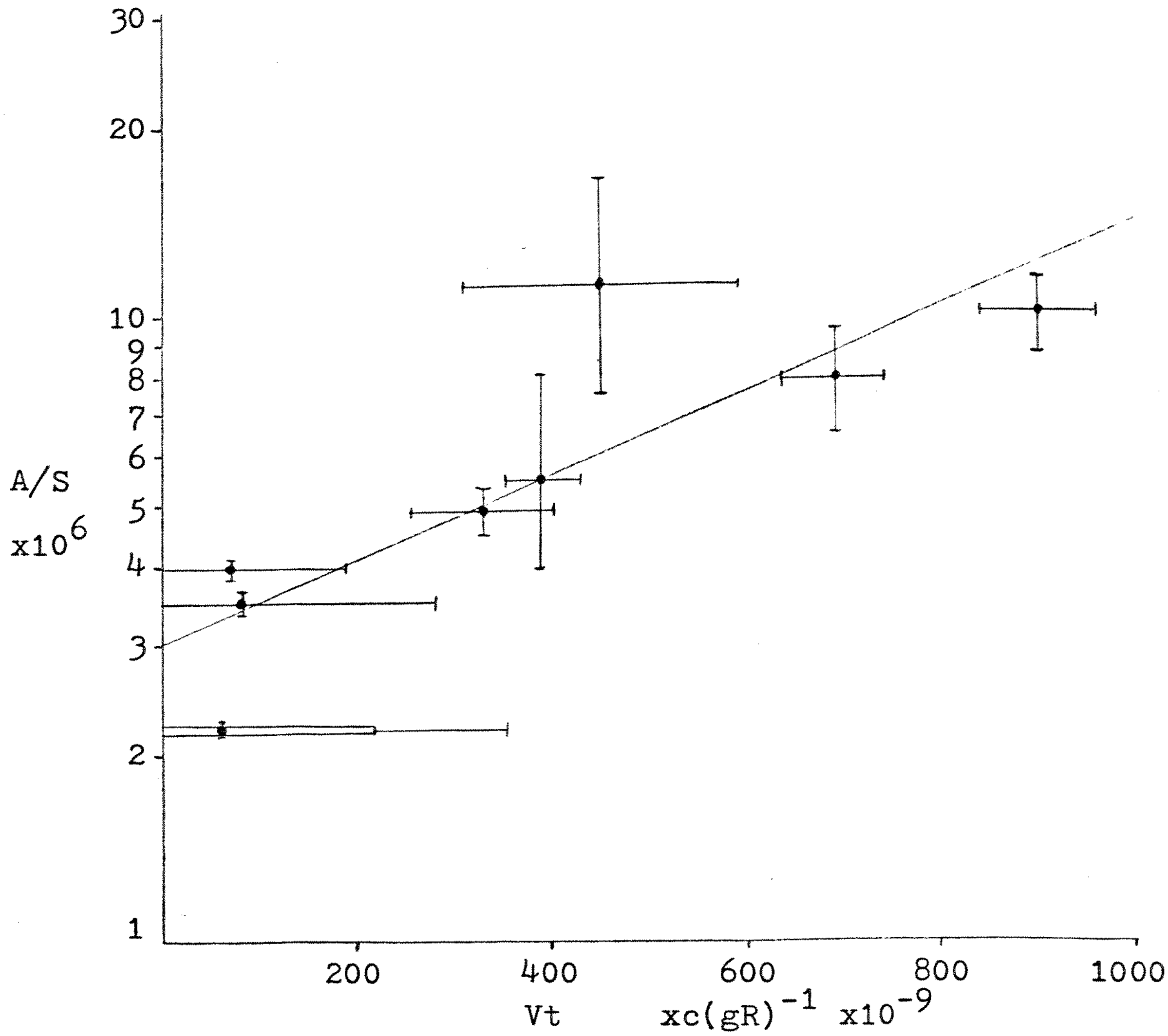


Figure 4

Semi-logarithmic plot of (Plate area/# hot spots) vs. (plate velocity x plate thickness). Horizontal error bars are errors associated with each plate from velocity and thickness. Vertical error bars represent ±1 hot spot. The double bar for Africa is the range of Vt, as its pole of rotation lies within the plate. The line was plotted using a least squares fit, with intercept 3.0×10^6, correlation coefficient .84, and certainty of correlation 99% (Young, 1962).

To be able to observe any of these limiting cases would be an ideal test of equation 1. Clearly, $V \to 0$ would be the most likely possibility, and the African plate is the most nearly stationary. In fig. 4, this would correspond to the y-intercept, at 3.0×10^6 km^2.

Exceptional Cases

Africa- The African plate's rotation pole lies within the plate. This causes large variations in velocity within the plate. These variations are represented by the double line in fig. 4. If the African plate was completely stationary, it would plot as a point on the y-axis, close to the intercept of the line described by all of the other plates. In fact, the African plate plots below the intercept and therefore apparently has excess hot spots. Most of the excess could be eliminated if all the hot spots that make up the Cameroon line had been counted as one. The Cameroon line is a linear sequence of hot spots trending northwest through Cameroon, in western Africa. Rather than being separate hot spots, so close together, they may represent the surface expression of one hot line (Richter and Parsons, 1975; Bonatti et al., 1977) as they trend in the direction of the last movement of Africa, as evidenced by the Walvis Ridge, but do not exhibit a clear age progression.

Regional Variations of Velocity and Thickness

Vulnerability Parameter- Pollack et al. (in press) related hot spot density to plate velocity and thickness in a way which allowed for the regional variation of V and T. They proposed a vulnerability parameter v.p.=$(u/kd)^{\frac{1}{2}}l$, where k and d are the thermal diffusivity and sublithospheric upwelling spacing, respectively (both considered constant), and u and l are the velocity and thickness of the lithosphere at any given point. The effect of v.p. on the hot spot distribution was represented in two bar graphs, one showing the percentage of the earth's surface in incremental vulnerability parameter ranges and the other showing the percentage of hot spots in area of incremental vulnerability parameter ranges.

The sublithospheric upwelling density cannot be determined from the vulnerability parameter defined by Pollack et al. (in press), because it goes to zero for a motionless or vanishingly thin plate. In trying to determine the sublithospheric upwelling density using a linear or power relationship between observed hot spots and plate characteristics, a singularity will develop if the velocity or thickness are zero, implying infinite upwellings per unit area. This is because in the relation between the vulnerability parameter and the sublithospheric upwelling density, the vulnerability parameter appears in the denominator, such that as v.p. approaches zero, the sublithospheric upwelling density goes to infinity.

If instead a vulnerability parameter is defined using an exponential relationship, the sublithospheric upwelling density can be determined (equation 1). There is no singularity since as V or T go to zero, the vulnerability parameter approaches 1. The influence of the lithosphere is thus effectively removed, and the sublithospheric upwelling density equals the hot spot density. Furthermore, the distribution so obtained is even more biased in favor of hot spots appearing in low vulnerability parameter areas (figs. 5 & 6).

Sublithospheric Upwelling Density

The results in fig. 4 show the sublithospheric area per upwelling to be 3.0×10^6 km^2, or 1700 km between adjacent upwellings. This sublithospheric upwelling density may have implications regarding the origin of hot spots and mantle convection, to be considered below.

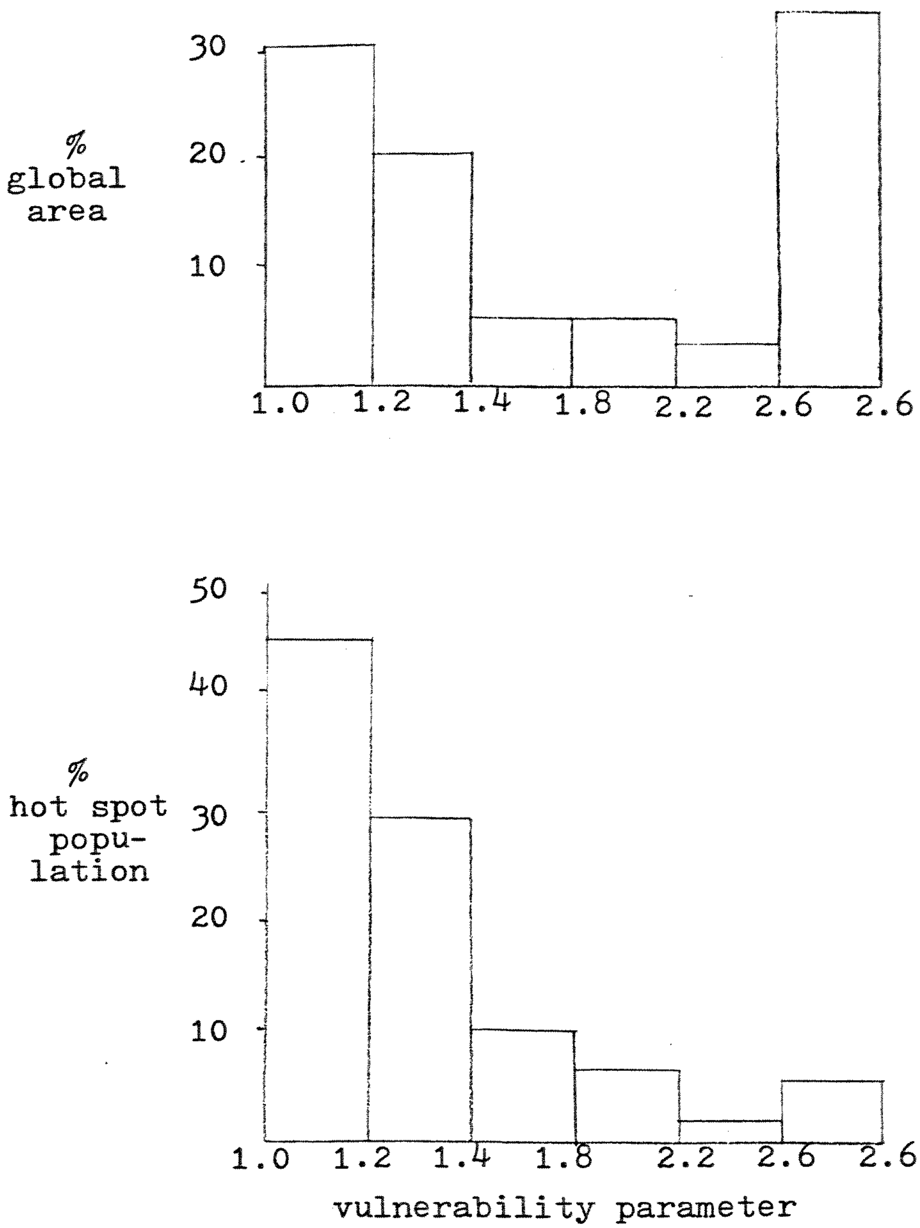


Figure 5

Bar graph of % global area and % hot spot population vs. exponential vulnerability parameter.

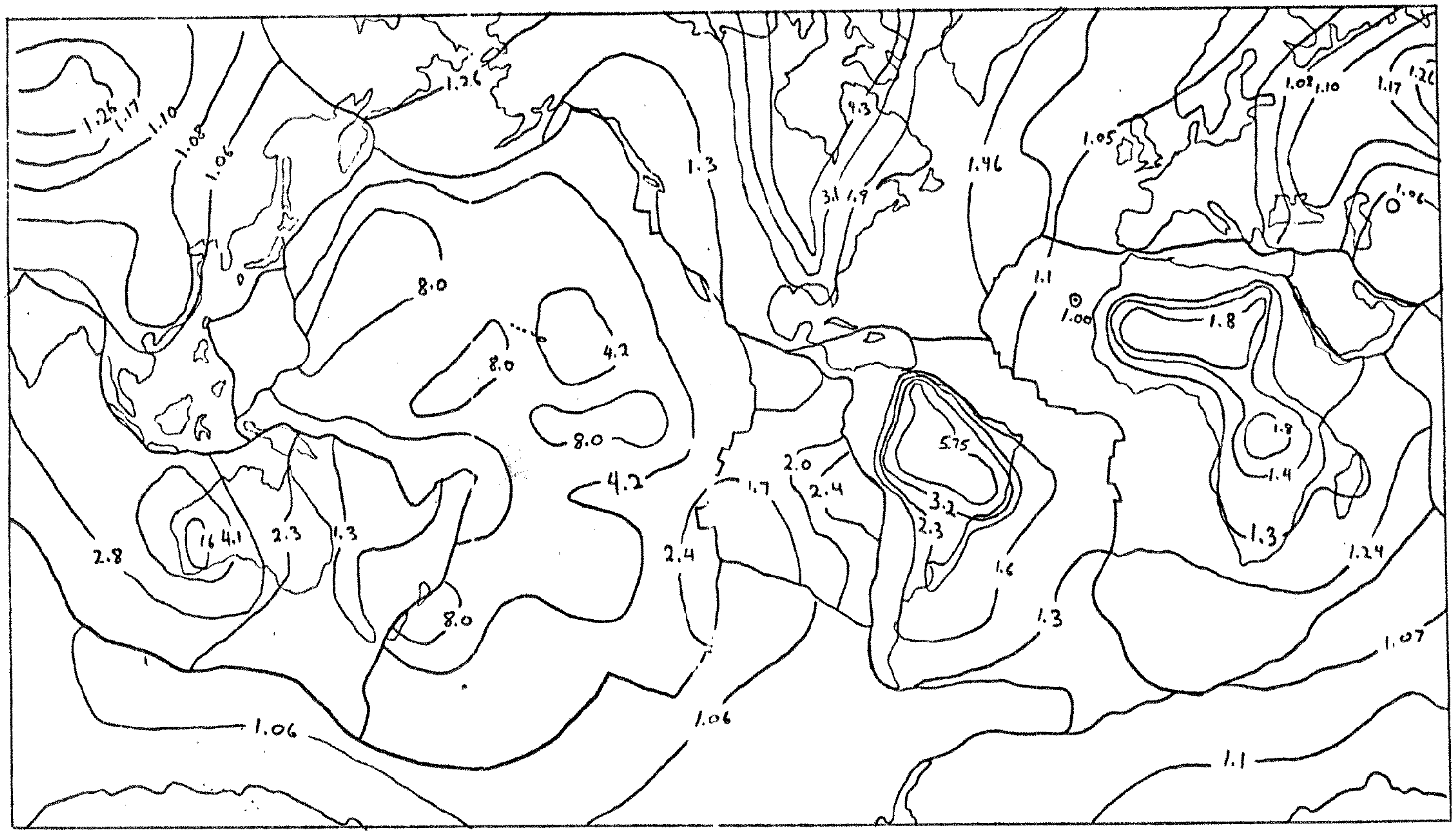


Figure 6
World map showing exponential vulnerability parameter

Contours represent exp(VT/R)

IMPLICATIONS OF UPWELLING DISTRIBUTION

The are several possible implications of the upwelling distribution regarding mantle convection. The depth of convection results directly from the upwelling density, and the scale of mantle convection can be inferred. The viscosity of the upper mantle follows from a constraint on the Rayleigh number.

Depth of Convection

The depth of convection, d, profoundly affects the nature and threshold of convection. For two or three dimensional convection patterns, if the wavelength, λ, is known, the depth of convection can be calculated from $\lambda=2^{3/2}d$ for three dimensional flow, and $\lambda=2d$ for two dimensional flow (Rayleigh, 1916; Oxburgh and Turcotte, 1978). The distance between upwellings (λ) was estimated above to be 1700 km. The corresponding depth of convecting fluid is about 600 km for three dimensional flow. By adding the thickness of the rigid lithosphere (120 km) the depth of convection becomes about 720 km.

650 km Discontinuity

There is evidence indicating convection to about 700 km. At about 650 km, there seems to be a seismic discontinuity which may confine convection to regions above this (Richter, 1977; Bonatti et al., 1977; Anderson, 1976; Grand and Helmberger, 1980; Muirhead and Hales, 1980;

Means and Jordan, 1980). Deep focus earthquakes are found at depths up to 720 km which establishes that convection occurs to this depth, since at this depth, lithosphere is still sinking into the mantle. Earthquake studies of Knopoff (1964) and Richter (1977) also show a sharp increase in penetrative resistance to subducting lithosphere at depths of about 650 km. There may be a phase transition (Anderson, 1976) and/or a chemical transition (Liu, 1979), at this depth, causing an increase in the density of the mantle.

The nature of the 650 km discontinuity is not clear. Refracted waves indicate that the discontinuity is spread out over a fairly large interval (200 km) (Helmberger and Engen, 1974), whereas reflected short period waves indicate a much sharper discontinuity (Engdahl and Flinn, 1969; Whitcomb and Anderson, 1970). The explanation may be found in the examination of the mineral assemblages and phase changes in this range of depth. Sammis (1976) considered the discontinuity to reflect a spinel-oxide transition. Anderson (1976) proposed the following scheme:

200-400 km	olivene + pyroxene + garnet
400-500	β-spinel + px-ga solid solution
500-670	$(\beta+\gamma)$-spinel + px-ga solid solution
670-800	γ-spinel + SiO_2(stishovite) <u>or</u> mixed oxides
800+	oxides <u>or</u> $(Mg,Fe)SiO_3$(perovskite) + MgO

He considered the nature of the reflection data to be a result of a px-ga solid solution to oxide transition, and the nature of the refraction to be a result of the β-γ spinel transition. Experiments by Liu (1977) show that Anderson's (1976) conclusions regarding the px-ga to oxides transition are erroneous. As an alternative, Liu (1979) proposed a chemical discontinuity across the 650 km boundary, as well as a phase transition. He suggested that in order for the 650 discontinuity to represent only a phase change, it would actually have to occur at some depth below 700 km.

An important density contrast exists across the 650 km discontinuity (Liu, 1979; Dziewonski et al., 1975; Hart and Anderson, 1977). This is where the mantle becomes more dense that the lithosphere, such that subducting lithospheric plates would 'float' on mantle material below the discontinuity (Richter, 1977; Anderson, 1976). Abarrier against further penetration would thus be presented to subducting lithosphere at about 650 km. The great seismic energy release found at a depth of about 700 km (Richter, 1977) further supports the existence of this barrier. The lithosphere can, however, penetrate the barrier to slight degrees (Means and Jordan, 1980), but it would be at this depth that the lithosphere would lose its negative bouancy (Anderson, 1976). Thus, the lithosphere would sink until about

700 km, where it would stabilize and float.

Isacks and Barazangi (1977) have asserted that deep seismic studies have shown that the lithosphere does not exhibit exclusively comressional stress at depth, but that it simply 'unbends', to again become parallel to the surface of the earth. This would create compressional stresses on the upper surface of the lithosphere, while creating tensional stresses on the lower surface. They (Isacks and Barazangi, 1977) use this data to refute any density contrasts at about 700 km, but neglect to consider the mechanism which 'unbends' the lithosphere. Coming to rest on top of a more dense material would seem plausible. In fact, this evidence would support rather than refute the density contrast at about 700 km (Oxburgh and Turcotte, 1978).

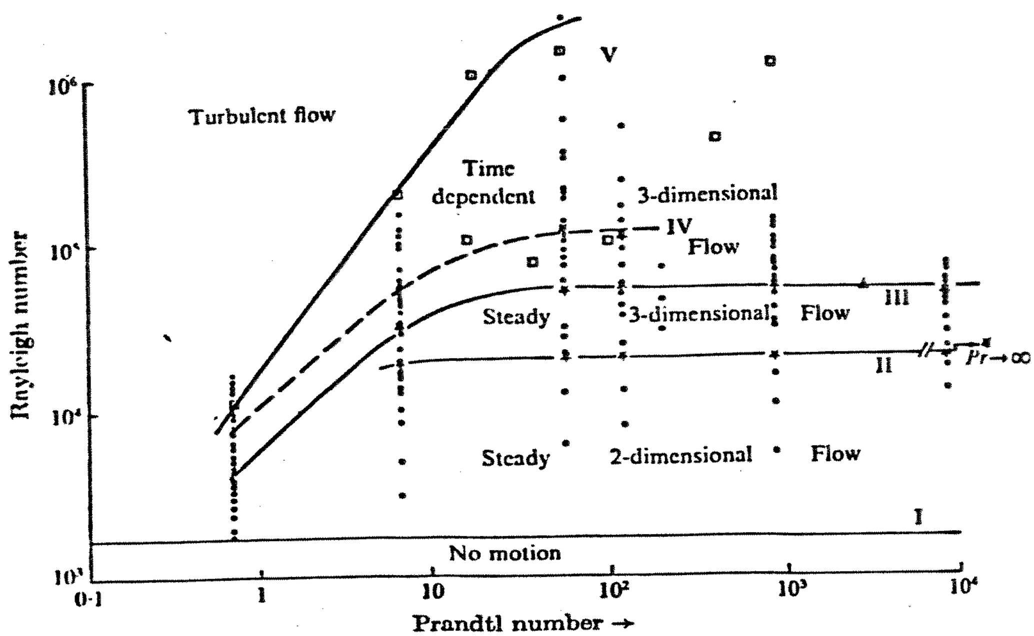
Parameters for Mantle Convection

If there is only a slight density contrast between layers in a stratified fluid, corresponding to small temperature differences of uniform material, the fluid will be stable, and any heat must be transported out by conduction. If the temperature difference is greater, however, convection will commence, and the fluid will flow. As the temperature difference is further increased, the flow pattern will change to one more efficient in transporting heat. The nature of the flow depends upon the Rayleigh number Ra, and the Prandtl number Pr where

$$Ra=\frac{\alpha\beta g d^4}{\nu k} \qquad Pr=\nu/k$$

where α=coefficient of thermal expansion, g=gravity, k=thermal diffusivity, β=thermal gradient, d=depth of convection (thickness of convecting fluid), and ν=viscosity.

The Prandtl number can be considered infinite for mantle materials since the viscosity is very large and the thermal diffusivity is very small. For small Rayleigh number, there will be no flow, and conduction will be the dominant thermal process. At the critical value of Ra, Ra_c=1708 (Knopoff, 1964; Oxburgh and Turcotte, 1978), flow will commence as a series of two dimensional rolls. The geometry of these rolls will be that of adjacent cylinders rotating in alternating directions. At higher Ra, Ra=13Ra_c (Krishnamurti, 1970; 1973), a transition takes place from the two dimensional rolls to a three dimensional geometry of rising plumes. This three dimensional aspect first appears as travelling waves moving along the tops of two dimensional rolls (Lipps, 1976). As Ra increases, the amplitude of the waves becomes greater, and the two dimensional character is lost completely. At higher Ra, there are other transitions to three dimensional time-dependent flow, and finally, turbulent flow (Krishnamurti, 1970; 1973) (fig. 7).



Turbulent flow
Time dependent
V
3-dimensional
IV
Flow
III
Steady
3-dimensional
Flow
II
Pr→∞
Steady
2-dimensional
Flow
I
No motion
Rayleigh number
10⁶
10⁵
10⁴
10³
0·1
1
10
10²
10³
10⁴
Prandtl number →

Figure 7

Viscosity

The spacing of sublithospheric upwellings may indirectly constrain mantle viscosity as a result of the determination of the depth of convection. For a system of convection with a base at about 720 km, the maximum viscosity that will allow convection can be estimated from the definition of Ra, such that

$$\nu = \frac{\alpha \beta g d^4}{Ra\ k}.$$

The following values are used in the calculation: Ra=1708, d=600 km, g=980 cm/sec^2, α=2x10^{-5} $^oC^{-1}$, β=10^{-5} ^oC/cm, and k=10^{-2} cal/cm^2-sec. With these values, the estimated maximum viscosity of the mantle, to a depth of 720 km, is 1.5x10^{23} poises. At this viscosity, two dimensional rolls will appear. If the viscosity is one thirteenth of this, or 1.2x10^{22}, three dimensional plumes will form.

Mode of Mantle Convection

There is evidence that suggests that hot spots are the result of both two dimensional and three dimensional convection. Hot spots like Hawaii and Yellowstone seem to be radially symmetric plumes resulting from three dimensional convection. The Cameroon line and Easter Island chain however, seem to be the result of two dimensional flow (Bonatti et al., 1977) since there is not a clear age progression in the direction of movement of the plates. Rather, some simultaneous volcanism is observed along greater linear extents than would normally be considered to be compatible with the radius of a plume. If both types of convection exist in the mantle, the viscosity of the mantle should not be very different from the value corresponding to the transition from two dimensional to three dimensional flow, 1.2×10^{22} poises, as estimated above. In places where the upper mantle is less viscous (or slightly thicker), the convective zone produces three dimensional plumes, In more viscous or thinner places, two dimensional rolls are produced.

<u>Review of Viscosity Literature</u>

The most common established technique for estimating mantle viscosity has been the consideration of isostatic adjustment of glacial depressions. Some (Artyushkov, 1971; Walcott, 1972a,b; Daly, 1980) have considered glacial loading and rebound as strictly aesthenospheric effects. However, this mechanism should cause a peripheral bulge around the area of ice loading. The bulge results from squeezing out a layer of about 100 km (aesthenosphere) and would now be subsiding by this type of model. The geodetic data, however, call for a sympathetic depression of immediately adjacent areas during loading, and uplift of these areas during rebound. This type of model also cannot acount for the negative gravity anomalies in Fennoscandia, indicating about 200 m of uplift remaining before isostatic equilibrium is attained. This then, calls for the participation of deeper material in glacial rebound. In order to account for all of the observed data, some (Cathles, 1975; Peltier, 1979; Post and Griggs, 1973) have considered flow not only in the aesthenosphere, but in the rest of the upper mantle as well. These considerations have accounted for the peripheral effects and amount of uplift still remaining, and have led to a viscosity of about 10^{20} poises for the aesthenosphere, and about 10^{22} poises for the rest of the upper mantle. McConnell (1968) developed a model which calls for a low viscosity layer with a viscosity of about 4.1×10^{21} poises just below the lithosphere, decreasing to 2.7×10^{21} poises between depths of 220 and 400 km, then increasing to at least 6.8×10^{22} poises below 1200 km and probably much higher (10^{25}).

Dicke (1969) calculated the relaxation time for second order distortions of the earth to be 870-1600 yrs, implying a mantle viscosity of about 10^{22}, agreeing with uplift consideration. With this low a viscosity, he argues, deep mantle convection is required to maintain the equitorial bulge. Others (Schubert and Young, 1976; Sammis, 1976, 1977) argue that if the viscosity of the mantle were less than 10^{24} and if most (90%) of radiogenic heat originated in the mantle (less than 10% from the core), then heat transport would proceed so rapidly that the core-mantle interface would have cooled to temperatures below that of the melting point of iron, eliminating the liquid part of the core by now. It would seem then, from all of the above considerations, that the viscosity of the mantle increases from about 10^{20} poises in the aesthenosphere, to 10^{22} poises in the upper mantle, and possibly higher in the lower mantle.

Discussion: Scale of Convection in the Mantle

The scale of mantle convection has been the subject of considerable controversy in recent years, with different researchers taking various stands based on scant data and very loose constraints. The possibilities regarding scale are well defined, however.

1. Convection could be confined to the aesthenosphere with stability existing in the rest of the mantle.
2. Convection could be confined to the upper mantle (700 km) with a stable lower mantle.
3. Convection could be mantle-wide with cells extending from the core-mantle interface to the lithosphere.
4. There could be two separate systems of convection, one above, and one below 700 km.
5. Overstability could exist with standing waves at the interface of the upper and lower mantle.

The last three of these possibilities are listed by Richter and Johnson (1974). The first two are neglected since most authors agree that lower mantle convection is highly likely (Turcotte, 1980) due to its very high Ra of about 10^6. Likewise, overstability is not generally seriously suggested as a state for the mantle considering the available data.

Aesthenospheric Convection

There is evidence that the viscosity of the aesthenosphere is markedly lower than that of the rest of the mantle from the attenuation of seismic waves (Cathles, 1975). Taking 100 km as the thickness of this layer, the wavelength is 200 km for two dimensional convection, and 280 km for three dimensional convection. The Ra for the threshold of convection leads to a viscosity of 1.1×10^{19} poises as an upper limit. 9×10^{17} poises would be the upper limit for the viscosity in order to get three dimensional plumes. However, studies of isostatic rebound in Fennoscandia have indicated a viscosity of about 10^{20} poises, an order of magnitude greater than that which would allow any convection. Furthermore, even if the viscosity of the aesthenosphere was such that convection was possible, the wavelength would be such that upwelling separation distances would be of the same order of lateral extent as that of hot spots on the surface. Such closely spaced upwellings would seem like a continuum to the lithosphererather than discrete heating centers. Also, were there to exist upwellings on this scale, they probably would be incapable of penetrating the lithosphere due to insufficient heat (as opposed to the greater amounts of heat inherent in upwellings from greater depths), and these small upwellings would simply be quenched.

Whole Mantle Convection

Many (Dicke, 1969; Elsasser et al., 1979; Davies, 1979; O'Connell, 1977; Peltier, 1979) have preferred whole mantle convection as the most likely mode of mantle convection. Some of the more convincing arguments are as follows: Davies (1979) states that to confine 'return flow' (to replace that of the moving lithospheric plates and to conserve mass) to the aesthenosphere, the difference in the viscosity of the aesthenosphere and the rest of the mantle must be greater than 10^4 poises. To exclude the lower mantle from flow entrained by the plates, the difference between upper and lower mantle viscosity must be at least 10^3 poises. Further, to confine convection strictily to the upper mantle, the viscosity difference between upper and lower mantles must be at least 10^4 poises. Considering these viscosity gradients as unreasonably large, Davies (1979) concludes that the mantle must be convecting as a whole. Elsasser et al. (1979) argue that since the lower mantle is likely to be convecting, and since the convection cells must be 'equidimensional' in order to minimize viscous dissipation according to the Helmholtz theorum (Batchelor, 1973), the convection cells must extend from the base of the mantle to the lithosphere.

In this case, in order to get threshold convection, the viscosity would have to be less than 5.2×10^{28}, a very high value. This agrees with an upper limit of 10^{27} placed on lower mantle viscosity by Schubert and Young (1976). However, studies indicate the viscosity to be closer to 10^{22} to 10^{24} poises (Watt, 1975; Cathles, 1975; Schubert, 1976). If this is the case, the corresponding Ra is about 10^6, indicating time varying convection with a wavelength of about 8000 km. The discontinuity at 700 km would be no obstacle to such a mode of convection if the density contrast across the boundary were small (less than .1%), and deep mantle melts would be observed at the surface. Convection on this scale could not be approximated as planar, but would have to be considered in the context of a spherical shell (Young, 1979). The temporal variation of convection on this scale would manifest itself as individual blobs of rising material (Ramberg, 1972) rather than a steady flow. These rising blobs would extract heat from the core, causing the core-mantle interface to cool locally upon departure of a blob (Vogt, 1975; Jones, 1977).

Two System Mantle Convection

The sublithospheric upwelling density derived here implies a depth of convection corresponding to the upper mantle. It does not, however, exclude the likely possibility of the existence of convection in the lower mantle.

Review of Literature- Convincing arguments are to be found supporting the idea of two separate systems of convection, with one in the upper, and the other in the lower mantle. In this model, the strongest argument against whole mantle convection is the very existence of a liquid part of the core (Schubert and Young, 1976; Sammis et al., 1976; 1977). If the mantle were to be convecting in one cell throughout its depth, the transport of heat would be so efficient that the core-mantle boundary would by now be at a lower temperature than the melting point of iron, unless the viscosity of the lower mantle were greater than 10^{24} poises.

Probably the clearest presentation of the problem of modes of mantle convection is proposed by Richter and Johnson (1974) and Sammis (1977). Here the Rayleigh number is defined in terms of the thermal gradient, as it is normally, and also in terms of density such that

$$R_\rho = \frac{gd^3}{\nu k} (\rho_2 - \rho_1/\rho_1).$$

It is the relationship between Ra and R that will determine the mode of mantle convection (fig. 8). Sammis (1976) goes further to break down the transition of whole mantle to separate convection systems according to the relationship of $\Delta\rho$ to viscosity (fig. 9). It can be seen from the figures (8,9) that less than a .1%

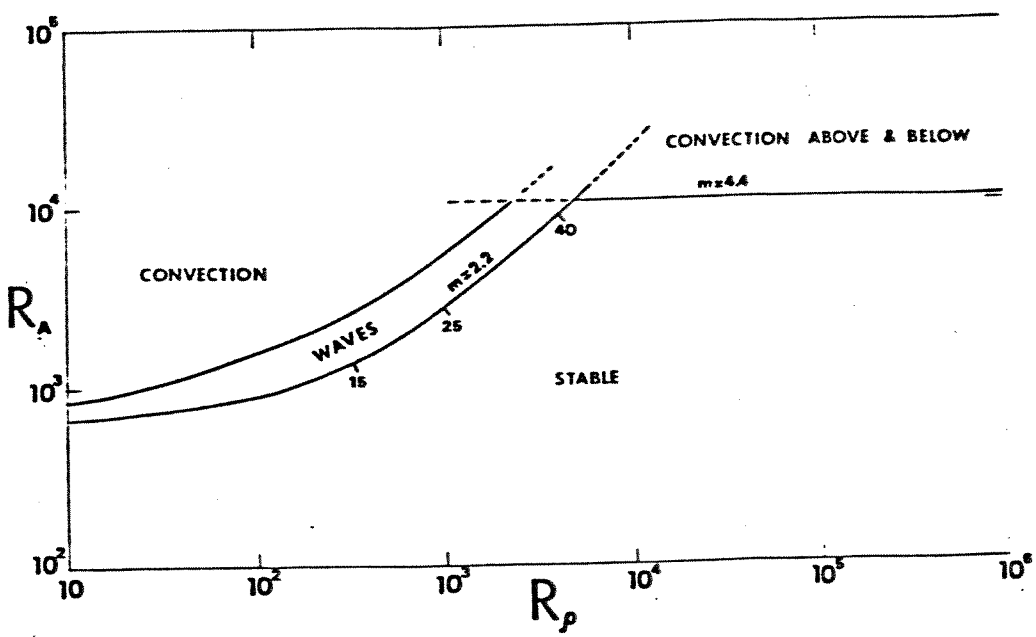


Figure 8

Plot of Ra vs R depicting modes of convection.
From Richter and Johnson, (1974).

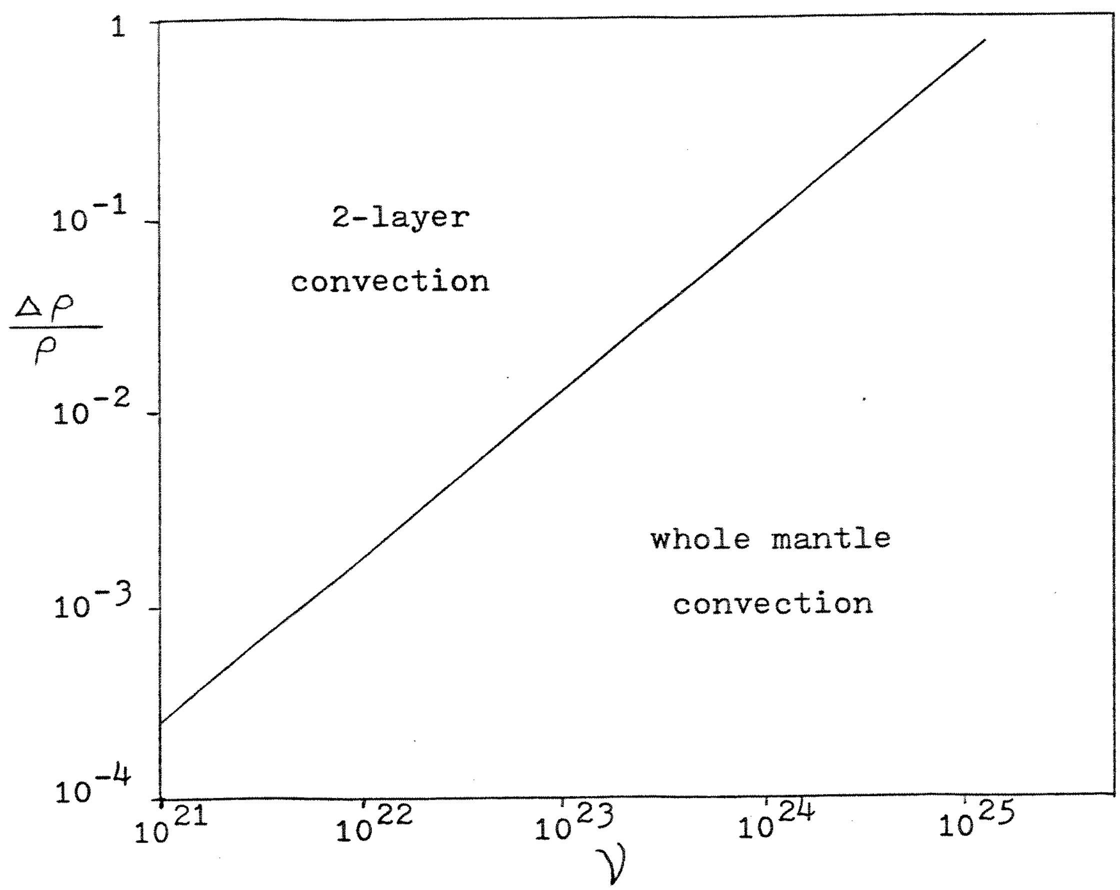


Figure 9

Plot of density difference across discontinuity vs viscosity, depicting modes of convection. From Sammis, (1977).

density contrast across the transition between the two layers will cause there to be separate convection systems. Clearly, the phase transitions and/or chemical transitions at about 700 involve density differences of much greater than this (Oxburgh and Turcotte, 1978; Shankland and Brown, 1980). It seems then, from present data, that the mantle is undergoing a two layer convection pattern.

Sun (1975) considers a somewhat different point of view. He agrees that at present, there is no mantle-wide convection, but he observes from studies of primary alkali basalts and nephelenites, that the alkali basalt source of his studies was deeper than present mantle convection. From this, he concludes that convection once involved a single whole mantle cell, but that at present, only the upper mantle is contributing magma to the surface of the earth. This agrees with petrologic studies of Hawaii by Parmentier et al. (1975).

Interaction between two systems of convection- If there are two separate systems of convection in the mantle, their mutual interaction becomes an important question. Although the lower mantle convection pattern may be confined to depths below 700 km, its thermal effects may be felt at the earth's surface in the following manner: Heat is brought to the 700 km level from the lower mantle via upwelling material. At this point the density contrast is too great to upweard convection to continue, so that the primary mode of heat transfer across the 700 km boundary is conduction. This heating of the base of the upper mantle in turn creates another convection cell in the upper mantle, which creates a hot spot at the surface. This double boiler mechanism may be responsible for plate motion. As a deep mantle upwelling blob impinges on the 700 km boundary, a broad area of the bottom of the upper mantle becomes heated. Upper mantle upwellings created above this area, in combination with those from an adjacent area, may be capable of thinning the lithosphere enough to cause continental breakup, thus initiating mid-ocean ridge spreading. This may explain the difference in scale between mid-plate hot spots and mid-ocean ridges, the former being due to the upper mantle convection system, and the latter indirectly due to the lower mantle convection system. The lower mantle convection system may be responsible for hot spots found along mid-ocean ridges.

It seems than, that genetically, there may be two kinds of hot spots- intraplate hot spots, and mid-ocean ridge hot spots. Examination of mid-ocean ridge hot spots may be an important line of research in the investigation of plate tectonics.

CONCLUSIONS

The correlation of Vt and ln(A/S) for the major lithospheric plates implies a sublithospheric upwelling density of 3.0×10^6 km^2/upwelling. The appearance of hot spots above upwellings is inhibited by higher plate velocity and greater plate thickness.

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