THE INFLUENCE OF PLATE MOVEMENT ON THE EVOLUTION OF HYDROTHERMAL CONVECTION CELLS IN THE OCEANIC CRUST

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ABSTRACT

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Two types of hydrothermal circulation can be distinguished in the oceanic crust, a very intense convection at spreading centers caused by the intrusion processes at mid-ocean ridges, and a much less vigorous, but more common convection in older, sediment covered crust driven by heat coming from the underlying, cooling plate. We investigated the relation between these two types of convection under normal spreading conditions, i.e. oceanic plate moving away from a stationary spreading center. In particular, we studied the stability of convection cells in a moving plate, the transition between on- and off-axis convection and the temperature distribution in these convection cells. The study is based on numerical calculations using the theory of flow through porous media.

Our results show that convection cells associated with the intrusion processes in the accretion zone of a spreading center are stationary with respect to the ridge axis. Convective heat transport in the stationary, on-axis cells is sufficient to remove all the heat released at the ridge axis. The other convection cells, which are not immediately associated with these intrusion processes, are not stationary. Most of them, especially once they are more than 30 km away from the ridge axis, move with the moving plate. This movement occurs regardless of the permeability distribution in the crust. As a consequence, individual segments of the crust are exposed either to continuous upwelling or to continuous downwelling flow once they have left the vicinity of the ridge axis; high water-to-rock ratios can be reached in these long-lived cells in spite of the relatively slow flow velocities. Temperatures in the off-axis cells, even in close proximity to the ridge axis, are low, in general below 100°C. The low-temperature alteration found commonly in the oceanic crust is evidence for the widespread occurrence of these off-axis cells. More specifically, the distinct differences in degree of alteration observed in some closely spaced DSDP cores are in good agreement with the concept of convection cells attached to the moving plate.

Because of the movement of the off-axis cells away from the stationary axial cells, a transition zone where new convection cells are formed exists at distances between 5 and 25 km from the spreading center. The formation of a new convection cell is accompanied by a maximum in low-temperature upwelling flow. When the heat transport by the plate and that by fluid convection is of similar magnitude, off-axis cells shift position with respect to the ridge axis and the moving plate. During this stage two cells can merge, which event causes a short (ca. 5000 years) episode of intense upwelling. These episodes, which are not directly related to any intrusive activity, can occur several times in or near a particular segment of the crust. Only during these short upwelling episodes are temperatures up to 200°C

reached in the off-axis cells. Chains of hydrothermal mounds near active spreading centers as well as low-temperature hydrothermal deposits such as the Mn lenses commonly associated with ophiolite sequences may be related to the formation of these off-axis convection cells.

INTRODUCTION

Hydrothermal activity associated with the spreading process at mid-ocean ridges has recently attracted considerable attention. The discovery of active hydrothermal vents in places such as the Galapagos Spreading Center (Corliss et al., 1979) and the East Pacific Rise (Spiess et al., 1980) has profoundly altered our understanding of hydrothermal processes and their relation to the formation of ore deposits and the alteration of oceanic crust. Although all of these vent systems were found at the axes of active spreading centers, several lines of evidence indicate that convection is not restricted to the immediate vicinity of mid-ocean ridges. The pronounced discrepancy between observed and predicted heat flow found in oceanic crust up to 80 m.y. old (Anderson et al., 1977), the wave-like heat flow distribution observed in closely spaced heat flow measurements over crust as old as 55 m.y. (Anderson et al., 1979), and element distributions in pore waters and the surrounding sediments (Maris and Bender, 1982; Sayles and Jenkins, 1982) all suggest the presence of convection in older oceanic crust.

Although the convection in older crust certainly is less vigorous, it probably circulates a larger mass of water through the crust than the convection at active spreading centers. Estimates put the contribution of on-axis flow to the total convection of water through the oceanic crust to 20% or less (Fehn and Cathles, 1979a; Sleep et al., 1980). Off-axis convection might thus play as significant a role for the interaction between oceanic crust and the oceans as on-axis convection, although the lower temperatures combined with the higher water-to-rock ratios to be expected in the off-axis convection will result in a significantly different nature of this interaction (e.g. Mottl, 1983; Seyfried and Mottl, 1982).

From a physical point of view the two systems, on-axis and off-axis convection, reflect two different processes: Axial convection is caused by the presence of strong lateral temperature gradients associated with the newly intruded material on the ridge axis and, in some cases, may even reflect the presence of a magma chamber. The widespread occurrence of convection in older crust, in contrast, makes it unlikely that this convection is related to individual intrusion processes except for some special cases such as, for instance, hotspot seamounts (Horibe et al., 1983). Instead, convection in older crust is in general a Rayleigh–Benard type circulation driven by the heat flow from the underlying, cooling crust. Lister (1981) coined the expressions active (on-axis) and passive (off-axis) convections for these two types of convection.

Because the crust is cooling as it moves away from a spreading center, it can be expected that convection in general decreases in intensity with time until it stops
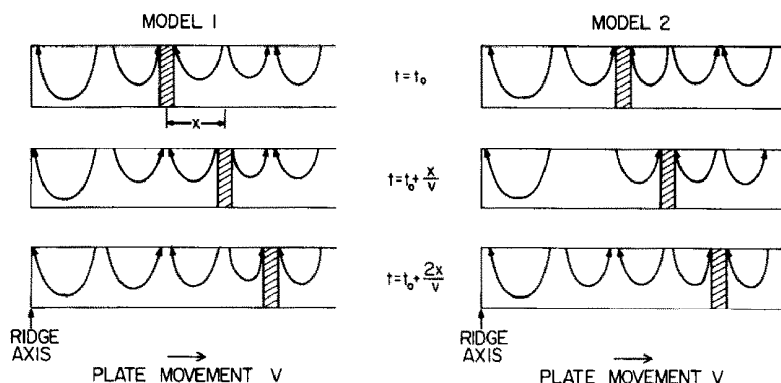
Fig. 1. Two models for the relation between on- and off-axis convection (see text for explanation). The position of convection cells and a segment of the crust (hatched area) is outlined for three times. In the lower two diagrams the plate has moved a distance of one (x) or two ($2x$) cell diameters away from the stationary spreading center.

completely. Consequently, it is plausible to assume that convection is a process which starts at the ridge axis and continues out to a distance where a combination of permeability decrease, temperature decrease and sediment cover prevents further convection. Except for isolated volcanic events, convection does not restart once it has been turned off. Thus, convection even in quite old crust is still related to the convection at the ridge axis. In this paper we will investigate the nature of this relation and, in particular, the behavior of hydrothermal convection cells in a moving plate.

The problem addressed in this paper can be posed in reference to two endmembers of possible convection configurations in the oceanic plate moving away from a stationary spreading center. In model 1 (Fig. 1), the plate moves through a continuous series of convection cells, initiated at the ridge axis. Because of the intrusion processes at the ridge axis, strong lateral temperature gradients exist, which cause the convection of seawater through the oceanic crust. The draw-down of cold seawater into these convection cells lowers the isotherms at some distance from the ridge axis and in turn causes a secondary cell to form. If the permeability in the oceanic crust is more or less uniform, this process could continue for long periods of time and cause the presence of a continuous array of convection cells. Although convection in older crust would be driven by the heat flow coming from the cooling plate, the arrangement of convection cells would still reflect the intrusion at the ridge axis even at great distances from the ridge axis. In this model a specific segment of the oceanic crust is exposed to alternating episodes of upwelling and downwelling flow as it moves through an arrangement of convection cells which is fixed with respect to the ridge axis.

In the second model convection cells not immediately adjacent to the ridge axis move with the moving plate as a consequence of the lateral heat transport associated

with the plate movement. In this case, individual segments of the crust are exposed either to continuous upwelling or to continuous downwelling flow once they have left the immediate vicinity of the ridge axis (Fig. 1). This model requires a transition zone where new secondary convection cells are generated when the distance between the axial cell and the first secondary cell becomes too large.

The numerical calculations reported in this paper indicate that off-axis cells move with the plate when the lateral heat transport by the plate is larger than the vertical heat advection by the convecting fluids. When the rate of fluid convection is small in comparison to the spreading rate, e.g., when the spreading rate is relatively large and the permeability is relatively small, all off-axis cells move with the plate. In the opposite case a transition zone exists within 5 to 25 km of the ridge axis where cells move with respect to the ridge axis and the oceanic plate. The area of particular interest for this investigation is thus the region between 5 and 25 km from the ridge axis where the transition between on-axis and off-axis convection takes place.

THE NUMERICAL MODEL

The numerical model is essentially the same as that used in previous publications (Cathles, 1977; Fehn and Cathles, 1979b; Fehn et al., 1983). It is described by the following three equations. Balance of momentum (Darcy's Law):

$$-\nabla p + \rho g - (\nu/k)q = 0 \tag{1}$$

balance of mass:

$$\nabla \cdot q = 0 \tag{2}$$

and balance of heat:

$$\rho_m c_m (\partial T/\partial t) = -\nabla \cdot q c_f T + K_m \nabla^2 T - \rho_m c_m v_s \cdot \nabla T \tag{3}$$

where c_f is the specific heat of the fluid (4186 J kg^{-1} deg^{-1}), c_m the specific heat of fluid saturated rock (837 J kg^{-1} deg^{-1}), g—gravitational field strength (9.8 m s^{-1}), k—permeability, K_m—thermal conductivity of fluid saturated rock (2.5 W deg^{-1} m^{-1}), p—pressure, q—mass flux, v_s—half spreading rate, ν—kinematic viscosity, ρ_m—density of fluid saturated rock (2700 kg m^{-3}), and ρ—density of fluid. We used the pressure- and temperature-dependent values of pure water, updated at each timestep, for density and viscosity of the circulating fluids in our calculations instead of the Boussinesque approximation.

Whereas the first two equations are standard for porous media models, the heat balance equation is adapted to the situation at mid-ocean. ridges by taking into account the heat transport caused by the movement of the plate away from a stationary spreading center. The change in the heat content of a segment of the crust depends in this case not only on the convective and conductive heat transfer but also on the heat transported by the moving oceanic crust, as expressed by the third term on the right-hand side in eqn. (3).

These equations are solved numerically by standard finite difference techniques (Carnahan et al., 1969). The emphasis of this investigation is on the role of the plate motion for the evolution of off-axis convection cells. For that purpose, we extended our computations to much longer periods of time than in our previous model calculations (Fehn and Cathles, 1979b; Fehn et al., 1983). Because we are here interested in the interaction between several convection cells, our domain of interest had to go significantly beyond the aspect ratio of 1 : 1 found normally in calculations of this kind. For our models we used a two-dimensional grid with 20 vertical and 80 horizontal grid points, equally spaced in both directions which covered a domain of 5 km depth and 31.2 km width.

Model parameters and boundary conditions

The standard model we chose for these calculations is shown in Fig. 2. The boundary at the ridge axis simulates a continuous intrusion process following Oldenburg's (1975) suggestion:

$$-K_{\mathrm{m}}(\partial T/\partial x) = \rho_{\mathrm{m}}v_{\mathrm{s}}\left[L + (T_{\mathrm{m}} - T_{\mathrm{f}}(t,z)c_{\mathrm{m}}\right] \tag{4}$$

where L is the latent heat of fusion (4.19×10^{5} J kg^{-1}), T_{m} the intrusion temperature (1200°C), and T_{f} the boundary temperature calculated at the end of the previous time step. The top (i.e. the seafloor) is kept at 0°C, and the right-hand boundary is insulating. In order to make the models comparable, a constant heat flow of 250 mW m^{-2} through the bottom of the domain was used in all the calculations, representing the heat coming from the cooling oceanic crust. The combination of the heat flow from the ridge axis and from the underlying plate results in a heat flow distribution similar to that predicted by plate models (e.g. Parsons and Sclater, 1977; Fehn and Cathles, 1979b). Free flow through the top boundary and no flow through the other three boundaries were chosen as flow conditions for the calculations.

In our models we assume a continuous intrusion process and a stable localization of the accretion zone. Studies of magnetic anomaly transitions at active spreading centers have shown that the accretion zone is quite narrow (between 1 and 10 km, depending on the spreading rate) and has remained localized for millions of years
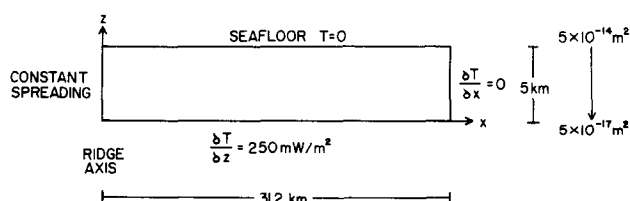

Fig. 2. The computational model.

(Klitgord et al., 1975; Macdonald, 1977; Macdonald et al., 1984) so that the latter assumption is probably applicable for most spreading centers.

The situation is different with regard to the nature of the intrusion process, which is probably episodic in reality. The times estimated for these episodes (Ballard and Van Andel, 1977; Macdonald, 1982) as well as the time necessary to cool an axial intrusion by convection (e.g. Lister, 1974; Cathles, 1983) range between 100 and 10,000 years. These times are shorter by at least one order of magnitude than the times characteristic for our models, as defined, for instance, by the time it takes for a segment of the crust to move the diameter of a convection cell. Although strong fluctuations in flow rates in the axial cells are to be expected as a consequence of individual intrusion episodes, the response time for convective heat transport is so short that all the extra heat introduced by individual intrusions can be removed within the actual accretion zone. For the formation of secondary cells it is then important only that convection exists at the ridge axis which is sufficient to remove all the excess heat associated with the intrusion process before the newly formed crust leaves the immediate vicinity of the ridge axis.

Our boundary condition for the ridge axis takes into consideration the total heat released over the depth of the domain (5 km) in accordance with the given spreading rate. It does not, however, accommodate the heat coming from a magma chamber at the ridge axis. Seismic investigations (Reid et al., 1977; Hale et al., 1982) as well as heat balance calculations based on the discharge observed from the vents at the East Pacific Rise (Macdonald et al., 1980; Converse et al., 1984) suggest the presence of a shallow magma chamber there. The latter calculations, however, also show that the convection through the presently active vents removes all the heat released by such a magma chamber within relatively short periods of time (< 40,000 years). Thus, even if a magma chamber is present, crust leaving the immediate vicinity of the ridge axis is essentially cooled because of the convection in the primary cell. Although the presence of a magma chamber will certainly cause much higher flow rates in the primary cells and, perhaps, also higher overall temperatures than in our models, the constellation of a dominant, but fixed primary cell and weaker, but moving secondary cells will not change. A stationary and continuous spreading process is thus an acceptable assumption for our purposes.

An important variable for our calculations is the permeability distribution in the oceanic crust. We chose here a permeability distribution which is homogeneous in lateral direction, but decreases exponentially by three orders of magnitude from a surface pemeability of 50 mD (50×10^{-15} m^2) over the depth of the domain. This permeability distribution is in good agreement with values assumed in previous calculations (e.g. Ribando et al., 1976; Fehn et al., 1983) as well as with the few in situ measurements of permeability in oceanic crust available so far (Anderson and Zoback, 1982). The penetration depth is based on the spacing of heat flow highs and lows near spreading centers (Davis and Lister, 1977; Green et al., 1981) and on the depth of alteration in ophiolite suites (Gregory and Taylor, 1981; Schiffman et al.,

1984). Closely spaced heat flow investigations near mid-ocean ridges (Williams et al., 1977; Davis et al., 1980; Green et al., 1981; Becker and Von Herzen, 1983a) and in older crust (Anderson et al., 1979) show a relatively smooth variation in heat flow and at most only a weak correlation with the presence of fractures. This observation suggests that inhomogeneities in permeability such as major fractures do in general not control the convection pattern in the oceanic crust, although they probably influence the local distribution of fluid flow. Since we are mainly interested in the behavior of convection cells in a moving plate, we are content here with a model with a reasonable permeability distribution, i.e. homogeneous in the horizontal and exponentially decreasing in the vertical direction, realizing, however, that there exist arguments (e.g. Johnson, 1979; Young and Cox, 1981; Becker et al., 1982) which suggest rather different permeability distributions.

For the same reason we neglect here influences which we consider secondary for our investigation. This includes a sediment cover of the crust and the local topography. Although both of these parameters have probably strong influence on the distribution of discharge and recharge areas on a local scale (1 km or less) heat flow surveys from areas such as the Galapagos Spreading Center (Green et al., 1981) and the Juan de Fuca Ridge (Davis and Lister, 1977) suggest that the overall convection pattern is determined by the permeability distribution in the crust itself. Topography might have a stronger influence in situations where prominent features such as seamounts are present which could act as chimneys for the upward flow (Davis et al., 1980). Because topographic features, however, also move with the plate, our argument that convection cells are attached to particular segments of the crust is still applicable in this case.

Our models are two-dimensional. Although the heat flow distributions found near active spreading centers (e.g., Green et al., 1981; Becker and Von Herzen, 1983a) are in general two-dimensional, there are also indications that distinct differences in activity exist along the strike of active spreading centers (e.g., Crane, 1979; Ballard et al., 1984). It can thus be expected that convection cells become established in a three-dimensional fashion rather than in the elongated rolls envisioned here. Even for a three-dimensional distribution of activity along a ridge axis, however, it still holds that the main activity associated with the intrusion processes occurs in a very narrow area along the ridge axis which causes the fixation of primary cells there. Because we are also dealing here with a situation where on-axis cells are stationary, our results are likely to be applicable in a qualitative sense to this situation as well, in particular with respect to the formation and movement of off-axis cells. If the flow is completely confined to fractures parallel to the ridge axis (e.g. Strens and Cann, 1982), our model is not applicable because interaction between adjoining convection cells is not possible in this case. Convection cells even of this kind have to follow, however, the plate motion together with the fractures they are associated with.

RESULTS

The main objective of these calculations was to study the influence of the plate movement on the evolution of convection cells in the oceanic crust. The only term in the main equations, (1)–(3), which reflects this movement is the last term in eqn. (3). We thus can establish a reference model by neglecting this term in the calculations. The result is shown in Fig. 3, calculated for a spreading rate of 3.5 cm/yr, which in this case, however, enters only via the boundary condition (4) for the ridge axis. Here and in the following diagrams isotherms (solid lines) and stream lines (broken lines) are given for several intervals after the time (t_0) at which the system had reached quasi-steady state, i.e. when overall temperatures and mass fluxes no longer changed significantly between time steps, although local changes still could occur.

The flow distribution in this example is dominated by a strong convection cell at the ridge axis followed by a regular pattern of weaker secondary convection cells. Upwelling zones are spaced approximately 7 km apart except for the somewhat larger distance between the axial cell and the first secondary cell. We also notice that temperatures beyond the axial cell are barely above 300°C at the bottom of the domain, and that convection probably does not reach temperatures much higher than 100°C in the secondary cells. Flow and temperature distributions for the three times are indistinguishable. Thus, a stationary distribution of convection cells results if the lateral heat transport associated with the plate movement is neglected.

The situation is significantly different, however, if we take this lateral heat transport into account (Fig. 4) for the same spreading rate of 3.5 cm/yr. Whereas
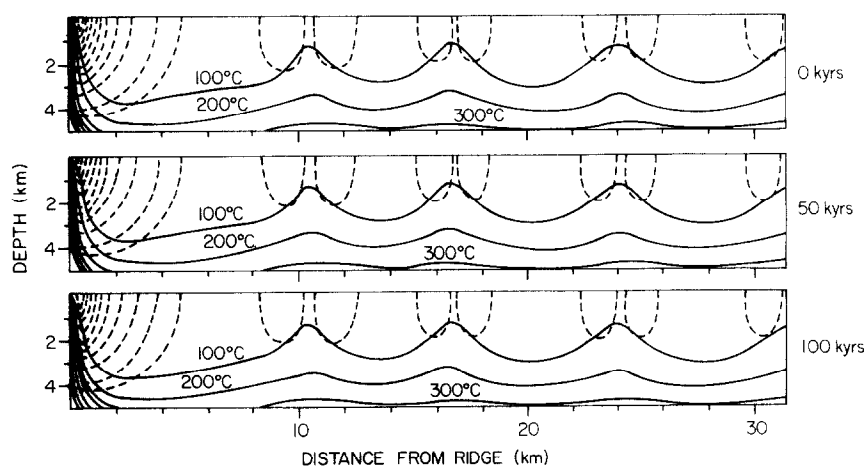

Fig. 3. Flow and temperature distribution for the standard model (spreading rate $v = 3.5$ cm/yr) without the spreading term in eqn. (3). Isotherms (solid lines, 100°C intervals) and stream lines (broken lines, intervals of 2.0) are shown at three different times after steady state has been reached.
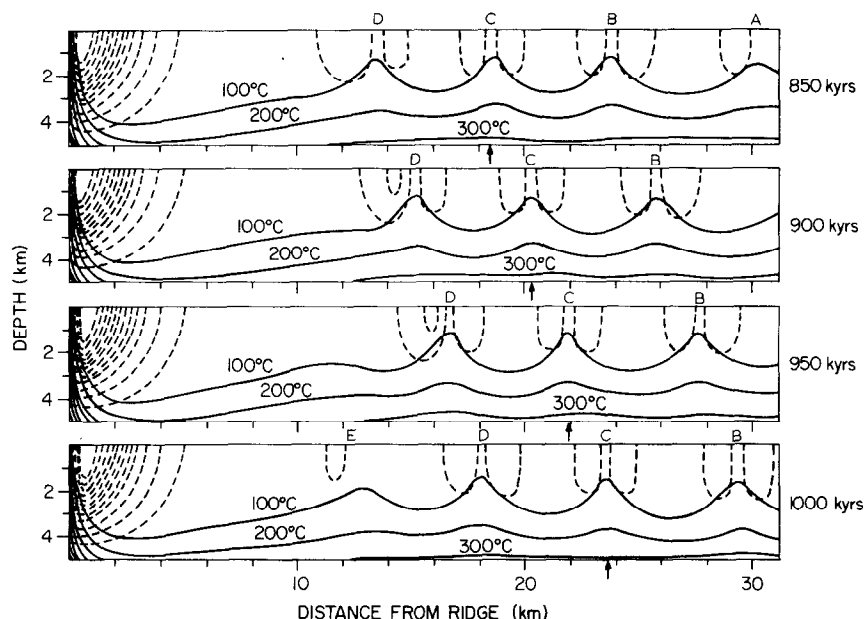
Fig. 4. Flow and temperature distribution for the same model as in Fig. 3, but with the spreading term in eqn. (3). The arrows indicate the position of a particular segment of the crust as it moves away from the ridge axis at a rate of $v = 3.5$ cm/yr. Letters $A - E$ indicate upwelling zones.

the axial convection cell is very similar to that in Fig. 3, the other upwelling zones, labeled $A-E$, are not stationary with respect to the ridge axis in this case. If we follow the position of a particular upwelling zone, we see that it moves away from the ridge axis. The arrows give the position of a particular segment of the crust at the indicated times; the movement of the upwelling zones follows closely that of the plate, i.e. the indicated segment is exposed to continuous upwelling flow in this case. As the convection cells move away from the ridge axis, the spacing between the axial cell and the first secondary cell becomes too large. Consequently, at 1 m.y. a new upwelling zone appears at a distance of 12 km from the ridge axis. The resulting distribution of convection cells at 1 m.y. years approaches that in the top diagram. Inspite of the movement of convection cells here, temperatures are very similar to those found in the previous case, i.e. circulating fluids reach temperatures above 200°C only in the axial cell.

The movement of the cells and the cyclical nature of this process become evident if we plot the position of upwelling zones with respect to the ridge axis as a function of time (Fig. 5). The dots, connected by lines, indicate the positions of the off-axis upwelling zones at particular times (for the sake of clarity locations are given at time intervals of 50,000 years; time steps in the calculations were between 125 and 250 years). The cells move away from the ridge axis, following closely the plate
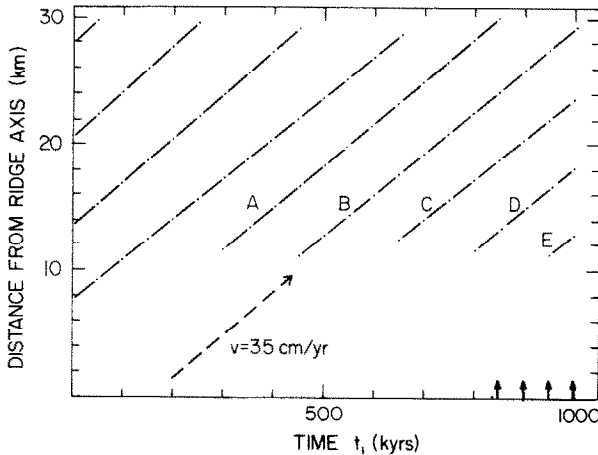
Fig. 5. Movement of the off-axis cells for a spreading rate $v = 3.5$ cm/yr. Each dot represents the position of an upwelling zone at time t_1 (calculated after quasi-steady state has been reached). The movement of these upwelling zones is parallel to the plate movement (broken line) at $v = 3.5$ cm/yr. The arrows indicate the four situations shown in Fig. 4.

movement at a rate of 3.5 cm/yr (broken line). As the distance between the ridge axis and the first upwelling zone increases, a new upwelling zone is formed at a distance between 10 and 12 km from the ridge axis. This pattern is repeated consistently in intervals of approximately 180,000 years, which correspond to the time necessary to move a segment of the crust over the distance between two upwelling zones.

The mass flux associated with these convection cells reflects a similar time dependence, demonstrated in Fig. 6, for the evolution of fluid flow with time as the five upwelling zones $A-E$ (see Fig. 5) move through the domain. The flow is integrated over the discharge area of each upwelling zone. Initially, the flow in the individual upwelling zones increases rapidly, goes then through a relatively broad maximum and finally reaches values around 1.3×10^5 kg yr^{-1} m^{-1} (ridge axis). All off-axis cells follow very similar patterns in this case.

The evolution of convection cells is quite different if we go to a slower spreading rate while leaving the other paramaters unchanged. Figure 7 shows four situations for a spreading rate of 1 cm/yr. Although the convection cells are also not stationary, their movement is more complex here than in the previous case. Whereas upwelling zones A and B, which are further removed from the ridge axis, follow the movement of the plate, zone C remains stationary for a while. As the distance between upwelling zones B and C becomes too large, two new zones (D and B_1) are formed. Once they are established, all zones move with the plate for a while before the process is repeated. Whereas temperatures in the axial cell are considerably lower here than those obtained for the two previous cases, temperatures in the secondary cells are quite similar in all three models.
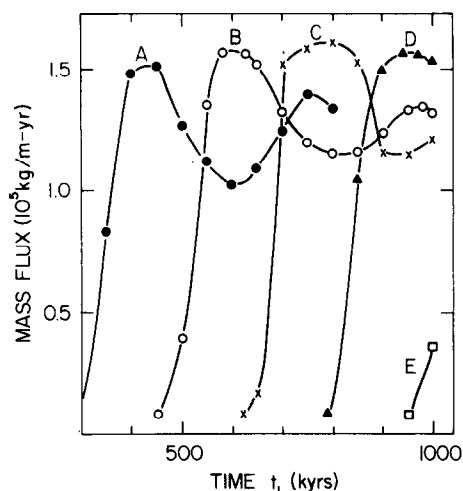
Fig. 6. Evolution of discharge rates for upwelling zones $A-E$ (see Fig. 5) with time t_1. The mass flux (in 10^5 kg yr^{-1} m^{-1} of ridge) is integrated over the width of each upwelling zone.
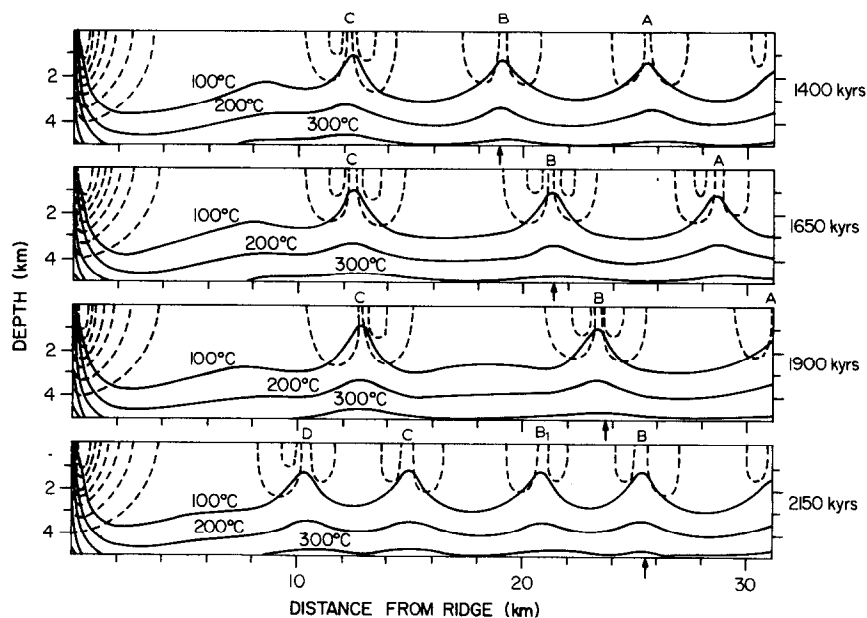

Fig. 7. Four episodes in the evolution of off-axis cells in crust spreading at a rate of 1 cm/yr. Arrows indicate the position of a particular segment of the crust at the given times. Isotherms and stream lines as in Fig. 3. Note the difference in the movement of upwelling zones A and B and that of upwelling zone C. Two new upwelling zones (D and B_1) appear at $t_1 = 2150$ kyrs.
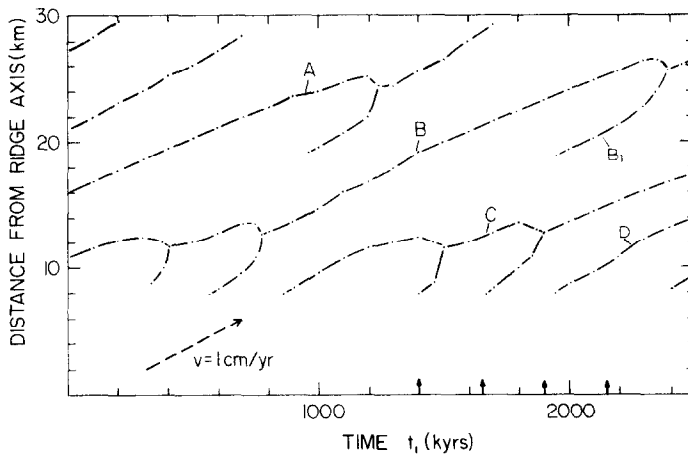
Fig. 8. Movement of upwelling zones at a spreading rate of $v = 1$ cm/yr. Dots mark the position of individual upwelling zones relative to the stationary ridge axis. The general movement of the upwelling zones is parallel to the plate movement (broken line), but is interrupted by merging episodes. The upwelling zones B and C undergo more than one merging episode in this example.

The movement of these upwelling zones is shown in Fig. 8. Again, we start with four upwelling zones which move with the plate. As a new zone is formed at 400,000 yrs, the next zone (B) does not move away fast enough and the two zones merge after approximately 100,000 years. Two of these merging episodes occur before zone B is far enough removed from the ridge axis to allow the formation of a new persistent upwelling zone C. Until this event, zone B moves less rapidly than the plate and thus changes its location both with respect to the plate and to the ridge axis. Because the upwelling zones move slower than the plate during these merging episodes, the distance between zones A and B becomes too large, and a new upwelling zone is formed between them. The whole process is repeated later between B and C: zone C also undergoes two merging episodes before it moves away from the ridge axis. Merging episodes occur also for zones A and B at distances of approximately 25 km from the ridge axis, corresponding to 2.5 m.y. old crust at this spreading rate.

These merging episodes are associated with intense upwelling as reflected in the flow evolution for the various upwelling zones (Fig. 9). Mass flux of zone B, for instance, starts out with values of somewhat over 10^5 kg yr^{-1} m^{-1} (of ridge), but then undergoes two short periods of much higher flow rates before coming back to the original rates. A third episode of intense upwelling occurs approximately 1.4 m.y. later. The other two upwelling zones follow quite similar patterns.

If we compare the temperature distributions in these models, we notice that they are quite similar, and, in particular, that high temperatures are reached only in the axial cell. Temperatures in the crust depend mainly on two parameters: the heat
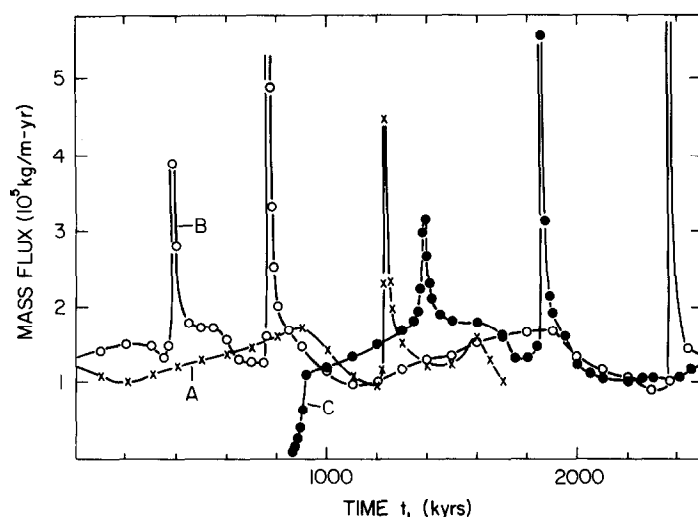
Fig. 9. Evolution of discharge rates (integrated over each individual upwelling zone) for upwelling zones A ($\times$), B (open circle) and C (closed circle) with time t_1 (time after quasi-steady state has been reached). The merging between two upwelling zones is accompanied by strong maxima in the discharge rates.

input from the ridge axis and from the cooling plate below, and the permeability distribution in the crust. Whereas the former parameter—as reflected in the boundary conditions—is reasonably well defined within the framework of plate tectonics (see, for example, discussions in Davis and Lister, 1974, and in Fehn et al., 1983) the permeability distribution is more open to discussion.

The dependence of the temperatures on the permeability distribution is evident if we compare models with surface pemeabilities ranging from 5 to 150 mD to the conductive solution (Fig. 10). As in the previous examples the permeability decreases exponentially over three orders of magnitude from the surface to the bottom of the domain. Because of the higher efficiency of convective than conductive heat transport, temperatures in the domain decrease considerably with increasing permeability. Most important for our discussion are, however, the temperatures reached by the convecting fluids. We can get a measure for these temperatures by comparing the position of the stream lines to the isotherms in these models. Although convecting fluids obviously penetrate deeper into the crust as the model permeability increases, a streamline such as the 1.0 streamline (which outlines the approximate extent of significant flow in the crust) does not cross the 200°C isotherms in these models except right near the ridge axis. In fact, the isotherms reflect quite closely the permeability at a given depth; in these cases, for instance, temperatures above 200°C do not occur in crust with permeabilities above 0.5 mD (except for the convection in the axial zone). Whereas the increase in permeability is accompanied by an increase of flow rates and depth of penetration and thus exposure of crust to hydrothermal alteration, temperatures reached in the convec-

$k_0 = 0$

100°C
200°C
300°C
400°C
500°C

$k_0 = 5 \times 10^{-15} \, m^2 \, (5mD)$ +1.0 −1.0 +1.0 −1.0 +1.0

−13.6 100°C 10^{-15}

200°C 10^{-16}

300°C

400°C 10^{-17}

$k_0 = 5 \times 10^{-14} \, m^2 \, (50 mD)$

−25.7 +4.8 −5.1 +4.4 −4.5 +4.5 −3.7 10^{-14}

100°C 10^{-15}

200°C

300°C 10^{-16}

$k_0 = 1.5 \times 10^{-13} \, m^2 \, (150mD)$ −3.7 +3.9 +3.3

−30.5 +6.3 −5.8 +4.9 −4.6 +4.7 −5.4 10^{-13}

10^{-14}

100°C

200°C 10^{-15}

DEPTH (km)

PERMEABILITY (m²)

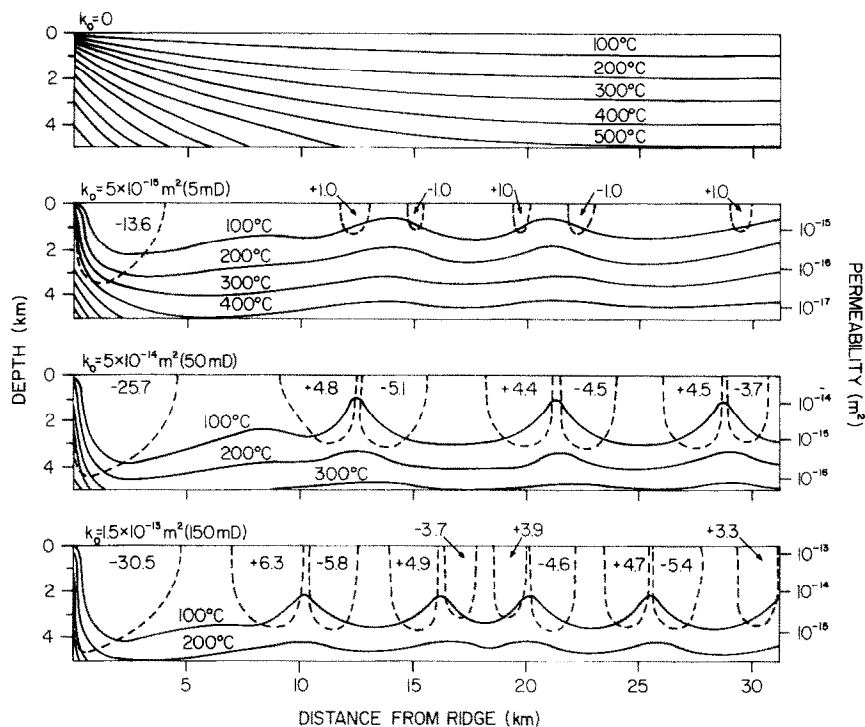5   10   15   20   25   30

DISTANCE FROM RIDGE (km)

Fig. 10. Isotherms and streamlines for three different permeability distributions compared to the temperature distribution in an impermeable crust ($v = 1$ cm/yr). Only the position of the 1.0 stream line is indicated as measure for the penetration depth of significant flow into the crust. Fluid flow penetrates to a depth where the permeability is approximately 10^{-15} m² and does not reach temperatures higher than 150°C in these models. The numbers in the convection cells indicate maxima or minima of the dimensionless stream function in the area enclosed by the stream lines.

tion cells are largely independent of the permeability in the oceanic crust for a given heat input. The observation that temperatures in the off-axis cells are quite similar throughout the domain also indicates that even at moderate permeabilities convective heat transport is sufficient to remove the excess heat coming from the ridge axis. The transition from on-axis to off-axis convection also marks the transition from high- to low-temperature convection near an active spreading center. These calculations thus indicate that convecting fluids in the oceanic crust do not reach temperatures above 200°C and that the bulk of the convection occurs at temperatures below 100°C, unless the convection is directly associated with volcanic intrusions as at the axis of a spreading center.

DISCUSSION

Three principal results were obtained in this study: (1) the movement of the off-axis cells with the plate, (2) the existence of a transition zone between on- and

off-axis convection, and (3) the restriction of the off-axis convection to low temperatures.

Movement of off-axis cells

The movement of the off-axis convection cells is caused by the lateral heat transport associated with the plate, which is sufficient to overcome the heat transport associated with the fluid flow within convection cells. This result was obtained for a permeability distribution uniform in lateral direction. Such a permeability distribution represents one endmember of the spectrum of possible permeability distributions in the oceanic crust, with the other endmember being a medium where flow is controlled completely by fractures. It is clear that model 1, i.e. convection cells stationary with respect to the ridge axis, requires a more or less uniform distribution of permeability in lateral direction in order to allow the plate to move through such an arrangement. A fracture-controlled distribution of convection cells, on the other hand, would favor the second model. The result that convection cells move with the plate even for a uniform permeability distribution in the crust allows the conclusion that convection cells are attached to a particular segment of the crust regardless of the permeability distribution in the crust. The exceptions are the convection cells directly associated with the intrusion process at the ridge axis, and some off-axis cells close to the ridge axis, where a combination of high flow rates and slow spreading rate causes movement of cells relative to plate and ridge axis. Individual segments of the crust are thus exposed to continuous upwelling or downwelling flow as soon as they have left the vicinity of an active spreading center.

Two main consequences follow from these considerations: The crust should show evidence for flow persistent in direction over long periods and for distinct differences in the degree of alteration on a scale of 5 to 15 km. There are numerous examples which can be cited in support of the former. Most of the holes drilled into the basaltic crust show alteration which can be attributed to persistent penetration of cold seawater into the crust. Examples are holes 332 and 335 (Robinson et al., 1977), holes 407, 408 and 410 (Floyd and Tarney, 1978), hole 395 (Lawrence and Drever, 1981) and hole 418A (Muehlenbachs, 1979). The reason for this bias is probably a combination of sampling procedure (more holes are drilled into well-sedimented basins than into basement highs which might act as local conduits for upward flow) and of the nature of convection which tends to concentrate upward flow more than downward flow because of the change in thermal transport properties of water with temperature (Fehn et al., 1983).

A well-studied example for persisting downwelling flow is hole 395, drilled into 14 m.y. old crust west of the Mid-Atlantic Ridge. On the basis of isotopic investigations of oxygen in calcite veins Lawrence and Drever (1981) concluded that these veins formed at temperatures consistently lower than those expected for crust

of this age, and that some of them might be very recent. In addition, the isotopic composition of sulfur and carbon in these veins suggests very high water-to-rock ratios during the convection. We have here a core where the same flow direction prevailed over a period perhaps as long as 14 m.y., resulting in consistent alteration and high water-to-rock ratios.

One of the few examples where persistent upwelling flow is demonstrated is hole 417A, drilled into 110 m.y. old crust at the southern end of the Bermuda Rise. Alteration temperatures for this hole have been estimated to range between 40° and 14°C with a slow, but observable decrease in time (Muehlenbachs, 1979; Lawrence, 1979; Alt and Honnorez, 1984). Hole 417D, only 450 m removed from 417A, shows the same range of temperatures, but a much lesser degree of alteration (Byerly and Sinton, 1979; Lawrence, 1979).

Only little information on the scale compatible with our models is available from DSDP cores so far. Holes 417A and 418A provide probably the most convincing evidence for a case where persistent upward and downward flow of seawater or seawater-derived fluids coexisted in close proximity for long periods. In contrast to hole 417A, the alteration products in hole 418A, at a distance of 6.5 km from hole 417A, suggest persistent cold water penetration into the crust, perhaps with temperatures slightly increasing with time (Muehlenbachs, 1979). Strontium isotope ratios measured in several of these cores suggest an upper limit of 10 m.y. for convective activity recharged by seawater (Hart and Staudigel, 1979) whereas mass balance considerations argue for somewhat longer times (Donnelly et al., 1979; Humphries et al., 1979). At the spreading rate of this area (1 cm/yr), the plate moved 100 km in 10 million years. Had the cells been fixed relative to the spreading center, the flow direction should have reversed several times in this core. The situation at holes 417 and 418 can thus be interpreted as a hydrothermal convection system which has been active continuously as it moved away from the ridge axis.

Observations of the variations in magnetic field strength perpendicular to the plate movement may also reflect the fixation of convection cells to the moving plate. Whitmarsh (1982) suggests that these variations are related to long-term alteration associated with low-temperature convection. If this interpretation is correct, it indicates that fluids in three-dimensional convection cells flow also in the same direction over long periods of time, i.e. the convection cells are attached to a particular segment of the moving crust.

Transition zones

The transition zones between off-axis and on-axis convection exist as a consequence of the movement of the secondary cells with the plate and of the fixation of the primary cells at the ridge axis. In these zones new secondary convection cells are formed once the spacing between the stationary on-axis cells and the moving off-axis cells has become too large for the stability of the cell distribution. We found two types of transition zones:

Type 1. When the heat transport associated with the moving crust is larger than that carried by the circulating fluids, cells are formed at regular time intervals within the transition zone and, once formed, move with the plate without further complications. The results found for the spreading rate of 3.5 cm/yr demonstrate this case. The transition zone in this case is quite narrow and well defined; upwelling zones are regularly spaced and decrease in intensity only moderately as they move away from the ridge axis.

Type 2. A more complex transition zone develops if heat transport rates by the plate and by the convecting fluids are similar as in the case of the 1 cm/yr spreading rate models. In this case some of the off-axis cells do not follow the plate movement immediately but do so only after the convective heat transport has become significantly smaller than that associated with the plate movement. Because of the slower movement of the convection cells, it is possible that a newly formed upwelling zone merges with the zone already in existence, creating a pulse of intense, episodic convection. As our example showed, it is even possible that more than one area exists where this process could occur.

This type of transition zone is characterized by the occurrence of short-lived pulses of intense flow during the merging of two upwelling zones. These pulses last approximately 5000 years, and have flow rates comparable to those found in the axial cells. Temperatures, however, are still significantly lower than in the axial cells, usually below 200°C. These episodes represent short periods of high-intensity, low-temperature flow in older crust (between 1 and 2.5 m.y. in our cases) which are not associated directly with an actual intrusion process.

The occurrence of the two types of transition zones is a function of the relative dominance of lateral heat transport (caused by the plate movement) or vertical heat transport (caused by convection). It depends, in other words, on spreading rates, heat input and permeability distribution in the crust. If, for instance, we lower the permeability by a factor of 10 in the model with 1 cm/yr spreading, a transition zone of type 1 results, whereas a model ($v_s = 3.5$ cm/yr) with higher flow rates caused by a higher heat input and higher permeability than used here resulted in a transition zone of type 2 (Fehn et al., 1983).

It is also clear that whereas the occurrence of type 1 zones again is independent of the permeability distribution in the crust, the merging of convection cells is not possible in a crust where the flow is dominated by individual, widely spaced fractures. These, together with the above considerations, suggest that, although a transition zone necessarily will be present in spreading crust with a stationary spreading center, this transition zone will be in most cases of type 1 and only under special circumstances of type 2.

The width of the transition zone is determined by the relative magnitude of plate and convective heat transport. Whereas a type 1 zone is very narrow and well-defined (about 1 km wide), the width of type 2 zones is much broader. Although it is possible that merging episodes occur even in distances greater than in our example,

the decrease in heat flow from the cooling plate and the resulting decrease in convective activity with increasing distance from the ridge axis limit the occurrence of these pulses practically to the area covered in our models.

The distribution and composition of the hydrothermal mounds and associated sediments at the Galapagos Spreading Center may illustrate the effects of the evolution of hydrothermal cells in a type 2 transition zone. The mounds are found in two distinct linear arrays, roughly parallel to the ridge axis, at distances of 23 and 28 km, respectively (Williams et al., 1979). Although water at elevated temperatures has been observed inside these mounds, no water egression has been reported. Heat flow measured in the field closer to the ridge axis is considerably higher than that in the other field, but both are much lower than the heat loss rate calculated for the active fields at the ridge axis (Green et al., 1981; Becker and Von Herzen, 1983b). The sequence found typically in the mound sediments begins with a layer of relatively unaltered sediments, followed by a layer of strongly altered sediments, and capped with manganese crust (Honnorez et al., 1981). The hydrothermal alteration of sediments is clearly restricted to the mounds, evidenced by the absence of altered sediments in cores recovered from drill sites outside the mound fields. Each of these mound fields can be interpreted as the result of the formation of a new upwelling zone which, once established, moves away from the spreading center, attached to a particular segment of the crust. The formation of the manganese crust probably is the result of recent merging events which provided the higher flow rates and temperatures necessary for the deposition of this material. With a spreading rate of 3.5 cm/yr in the Galapagos area, the distance between the two fields represents a time difference of approximately 150,000 years between the formation of these upwelling zones, and agrees well with our models. If these considerations are correct we expect that several more areas of hydrothermal mounds are to be found in quite regular intervals away from the ridge axis, covered by increasing layers of sediments.

The distribution of manganiferous lenses in the Franciscan Assemblage could also be the result of pulses of rapid low-temperature convection that are characteristic of the second type of transition zone. These deposits are found as small lenses associated with basalts but embedded in chert layers. Commonly, two or three of these lenses occur above each other, separated by 1 to 8 m of radiolarian chert. These observations as well as other lines of evidence suggest a marine origin for these deposits. According to this interpretation, these deposits were formed in separate pulses of hydrothermal activity close to, but not at an active spreading center (Crerar et al., 1982). Although it is possible that these deposits simply represent the pulses associated with individual off-axis intrusions, their widespread and quite regular occurrence makes this interpretation doubtful. If, on the other hand, a situation existed where spreading and convective heat transport were of the same magnitude, short pulses of hydrothermal activity could have been quite frequent during the merging episodes within the transition zone. As our models

show, a particular segment of the crust can be exposed to several of these pulses as it moves away from the ridge axis, which would explain the existence of several lenses at the same location separated by layers of sediments.

A still active example of the second type of off-axis activity could be the recently discovered mound zones in the Marianas Trough, a back-arc spreading center which opens at a rate of 2 cm/yr (Hussong and Uyeda, 1981). Two distinct zones of very high heat flow and perhaps even discharge of hydrothermal fluids have been found there, on 2 m.y. and on 5 m.y. old crust, respectively (Hobart et al., 1983). The wide spacing and the apparently high activity in these two areas make it likely that we are dealing here with a situation in accordance with a type 2 transition zone, where widely spaced zones of episodic upwelling occur caused by the merging of convection cells.

Type 1 transition zones produce localized deposits of hydrothermally altered sediments. Because low temperatures and flow rates are dominant in these situations, this type of transition zone would in general not be associated with the deposition of metals such as the above mentioned manganese deposits. The lower sequence of sediments in the Galapagos mounds as well as the umbers found associated with the Troodos massif (Robertson, 1975) probably are examples of the evolution of low-temperature convection cells without merging events.

Temperatures in off-axis convection cells

In our calculations we found that high-temperature convection, i.e. convection where fluids reach temperatures in excess of 200°C is essentially restricted to a very narrow zone at the ridge axis. Even at moderate permeabilities convective heat transport is sufficient to remove all the extra heat coming from intrusions at the ridge axis. Although these results were obtained for a continuous spreading process, they are—as argued earlier—also applicable to the more realistic case of individual spreading episodes.

The important point here is that crust formed at the ridge axis will be cooled quickly and—before leaving the immediate vicinity of the ridge axis—will have reached temperatures which will be relatively constant over long periods of time. The transition between high-temperature convection ($T > 200°C$ in convection cells) and low-temperature convection occurs thus within the first few kilometers from the ridge axis. This result is in good agreement with observations in many DSDP cores which usually show evidence for a very short-lived, high-temperature stage, followed by long-lived, low-temperature alteration (e.g. Bohlke et al., 1980; Alt and Honnorez, 1984). The rapid collapse of high-temperature convective systems at mid-ocean ridges, such as that at the Mid-Atlantic Ridge, 23°N (Gallinatti, 1984) where samples all dredged from within the rift valley show the complete transition from very high ($T > 400°C$) to moderate and low temperatures ($< 200°C$) also support the idea of a fast and complete transition from high to low temperature convection very close to the ridge axis.

CONCLUSIONS

Based on numerical calculations we have developed a model for the evolution of convection cells in a moving oceanic crust with a stationary spreading center. Fluid flow near the ridge axis is dominated by a strong convection cell which removes all the heat introduced into the crust at the ridge axis, including the heat released from magmatic intrusions and probably also from a magma chamber. This axial convection cell is stationary at the ridge axis, i.e. the location of this cell changes only little with respect to the ridge axis although it might show quite dramatic fluctuations in flow rates and temperatures. All the other convection cells, which are driven by the heat coming from the underlying cooling plate, are not stationary but move in general with the plate away from the ridge axis, i.e. these convection cells tend to be attached to individual segments of the crust. Alteration caused by these secondary cells can therefore be expected to be localized and to show the effects of continuous upwelling or continuous downwelling flow. These effects have been found in many drill cores where basement has been reached, most spectacularly in DSDP sites 417A and 418A.

Because the secondary cells are shielded from the high temperatures at the ridge axis by the primary cell, temperatures in these cells are considerably lower than in the axial cells, in general below 100°C. On the other hand, whereas the residence time of an individual segment of crust is relatively short in the region of primary convection, flow in the secondary cells can go on for millions of years, so that a low-temperature alteration with high water-to-rock ratios can be expected there as observed in DSDP 395.

The formation of new secondary convection cells takes place in a transition zone at distances between 5 and 25 km from the ridge axis. The nature of the transition zone depends on the relative magnitude of heat advection by the moving plate and the advection by fluid flow. If the spreading rate is moderately fast and the permeability of the plate not very high, new off-axis cells are formed at a given distance from the ridge axis in relatively constant time intervals. Distance and intervals depend on spreading rates and permeability distribution in the crust. During the formation stage higher flow rates and slightly higher temperatures than otherwise typical for these cells occur. Localized occurrences of hydrothermally altered sediments are the consequence of the evolution of simple off-axis convection cells that move with the moving plate.

In the opposite case, when the spreading rate is slow and the permeability is high, the transition zone is wider and the movement of convection cells in this zone is more complicated. Off-axis cells can move in this case relative to the ridge axis and to the plate. Strong, episodic upwelling of low-temperature hydrothermal fluids occurs when some of these cells merge in the transition zone. These episodes are examples for high-intensity flow independent of the direct influence of an intrusive process. Flow rates during these episodes are comparable to those occurring in the

axial cells, but temperatures are still considerably lower, in general not higher than 200°C. A particular segment of the crust can undergo several of these upwelling surges as it moves away from the ridge axis. Hydrothermal deposits such as the mounds at the Galapagos Spreading Center, manganiferous lenses in the Franciscan Assemblage and the widely spaced zones of active mound areas near the Marianas Trough may be related to episodic upwelling of this kind.

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