Passive margin development. A consequence of specific convection patterns in a variable viscosity upper mantle

Continental margins
Convection plumes
Viscosities
Plate tectonics

Marges continentales
Panaches de convection
Viscosités
Tectonique des plaques

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ABSTRACT

Simple thermodynamics calculations are applied to processes of a rising plume within an environment of an upward directed gradient of viscosity in the uppermost mantle. Four stages of the development of passive margins are described. Following an initial regional uplift strong lateral movements are induced at asthenospheric levels, stretching the overlying lithosphere. In the second stage plume material also enters the low viscosity lower crust, replacing continental material and forming a new, higher crust-mantle boundary by gravitational differentiation. The initial phase lag between the laterally intruding plume material and the rest of the lithosphere disappears once the continent is rifted and sea floor spreading is established (= stage 3). Finally, at stage 4, quiet subsidence, controlled by contraction, sedimentation, and possible phase changes in the lower crust, is established. The sequence described is in accordance with many observations of geology and geophysics. Rising plumes and sinking slabs are considered equivalent expressions of the heat engine earth which determines global tectonics.

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RÉSUMÉ

Évolution des marges passives. Une conséquence des régimes particuliers de convection dans un manteau supérieur à viscosité variable.

Le comportement d'un panache ascendant depuis l'asthénosphère vers un milieu à gradient de viscosité positif en direction de la surface, situé dans le manteau supérieur, est étudié à l'aide de formules thermodynamiques simples. On décrit quatre étapes de l'évolution des marges passives.
Après un soulèvement initial, d'importants mouvements latéraux apparaissent au niveau de l'asthénosphère, provoquant l'amincissement de la lithosphère sus-jacente. Dans la deuxième étape, le matériel du panache pénètre dans les niveaux à basse viscosité de la croûte continentale, remplaçant le matériel continental et formant une nouvelle limite croûte-manteau moins profonde, par différentiation gravitaire. La surface de changement de phase entre le matériel du panache introduit latéralement et le reste de la lithosphère disparaît une fois que la distension continentale a eu lieu et que l'accrétion océanique est établie (troisième étape). Enfin, au cours de la quatrième étape, s'instaure une subsidence régulière, contrôlée par la contraction thermique, la sédimentation et des changements de phase éventuels dans la croûte inférieure. Cette séquence est compatible avec de nombreuses observations géologiques et géophysiques. Les panaches ascendants et les plaques en cours d'enfoncement sont interprétés comme une manifestation équivalente du « moteur thermique terre » qui détermine la tectonique globale.

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INTRODUCTION

Among many geologists and geophysicists who investigate continental margins, there seems to be a common consensus about the two basic stages in the development of passive margins :

1) a period of stretching with crustal thinning, block-faulting, and rifting ;

2) a period of spreading connected with quiet subsidence and sedimentation at the margins.

In the present paper this concept is wholeheartedly accepted. In addition, the period 1 is investigated in more detail and is subdivided into an "uplift" stage and a "block faulting" stage. Much weight is attributed to these first stages because a statement such as "in the beginning there was a sudden stretching" which is often found in the modern literature seems unsatisfactory. Also, the period 2 is subdivided into two stages in order to explain some observations of a mature margin. In general, an attempt is made to explain these stages as physical expressions of convection patterns in an upper mantle with a viscosity gradient.

Only few investigations deal with the first period. Heiskanen and Vening-Meinesz (1958), Girdler (1964), and Artemjev and Artyushkov (1971) already associate the stretching of the crust with "diverging convective flows in the mantle", the rising part coming from some depth below the asthenosphere. Also, investigators who do not believe in deep seated convection (e.g. McKenzie, 1978), have to assume an initial period of "sudden stretching" in order to explain the subsequent subsidence of the margin, its thinning, and its present heat flow. McKenzie's model is adopted by Le Pichon and Sibuet (in press), for example, in order to successfully explain most features of the Armorican and Galician continental margins of the Northeast Atlantic. Instead of "sudden stretching" some workers postulate a "sudden heating event" for the beginning of rifting, as for instance Bott (1980). It seems that neither of these postulated "sudden" events provide a convincing explanation and certainly do not seem to be reasonable from the rheological point of view which will be stressed in the present paper. There is no doubt that the later stages of margin formation *may* develop from the initial conditions postulated by the above mentioned authors, and an agreement generally is reached for the stage of spreading and subsidence with the accumulation of (mostly shallow-water) sediments. This later stage is well explained by :

a) thermal cooling and contraction of the growing lithosphere (Parsons, Sclater, 1977 ; Watts, Ryan, 1976) ;

b) the load of sediments in connection with *a* (Bott, 1973 ; Sleep, 1971) ;

c) Phase transitions, for instance gabbro to eclogite, in connection with *a* and *b* (Spohn, Neugebauer, 1978 ; Neugebauer, Spohn, 1978).

The contribution from both *a* and *b* is estimated by Steckler and Watts (1978) for Atlantic-type margins. Here, and in many other papers, the effects mentioned may well explain the majority of observations. The explanation *c* involving deep crustal metamorphism provides an attractive additional force in connection with *a* and *b* and may induce a step in subsidence. From seismic observations, however, the Moho in most areas is a continuous boundary from the deep sea into the continental shelf. This boundary is mostly so sharp and shallow that it seems at least doubtful whether the gabbro-eclogite transition which requires greater depths and greater transition intervals, really plays a major role.

The generally shallow Moho below most shelf areas certainly cannot be explained by a pure subsidence process of the lithosphere. Either the observed shallow Moho is a consequence of the early period of stretching — with listric faults at depths (Murawski, 1976) and an old Moho (Steckler, Watts, 1978) — or/and the shallow Moho is a consequence of intruding magmas into the lower crust below the shelf, forming a new Moho at shallow depths by differentiation and crystallization. This process would be similar to that below the ridges meaning that the process of ridge formation continues *below* the continental areas adjacent to the juvenile rift. It will be explained in the present study as a consequence of the convection or plume model, plumes rise from depths below the asthenosphere into lithospheric levels. Their movement depends on models in which viscosity strongly increases upward from the upper part of the asthenosphere toward the uppermost mantle and from the base of the crust toward the upper part of the crust, leaving the lower crust as another — though smaller — zone of a viscosity minimum (Vetter, Meissner, 1979). It should be pointed out that the existence of a viscosity minimum in the lower crust is a function of the water content, the temperature, and the material of the lower crust (Meissner, 1974). As the temperature at the base of thick continental crusts of moderate age is generally higher than in old shield areas and as the material of the continental crust generally shows lower activation energies and lower melting points T_m than the peridotitic material of the mantle, most model calculations relating viscosity η to the ratio of T_m/T result in viscosity minima for the lower continental crust (Vetter, Meissner, 1979). There are many observations from reflection seismics and tectonic style which strongly support the assumption of such a low viscosity layer in the lower continental crust, and only some of them can be mentioned here : there are the "cushions" in continental rifts ; the horizontal, often laminated, reflectors in the lower crust which nearly always exhibit a strong increase in reflectivity compared to the poor reflections in the heterogeneous upper crust (Meissner, 1979). There are the "squeezed-out" lower crusts in areas of depression where the Moho is elevated and there are mountain roots often laterally displaced from the area of highest elevation (Prodehl, Pakiser, 1980 ; Meissner, Vetter, 1974). Figure 1 gives an estimate of the viscosity depth function after Vetter and Meissner (1979), which is used as a kind of starting model for the present study. A low-viscosity lower crust is certainly to be expected in the thick crusts of orogenic areas where Atlantic-type rifting began.

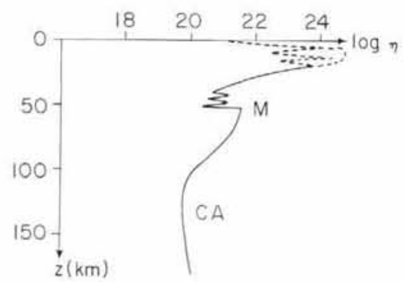

Figure 1

Viscosity-depth model of an orogenic area. M = Moho ; CA = Continental asthenosphere. Upper crust : heterogeneous ; lower crust : ductile with cristallisation seams.

THE PREPARATORY STAGE OF RIFTING AT ATLANTIC-TYPE MARGINS

There is no doubt that rift and spreading-type convection patterns were already established during the opening of the Pre-Atlantic, the Japetus ocean, which may have started as early as 1,000 m.y. ago. Other distant convection patterns, possibly associated with beginning subduction on both sides of the Japetus became, however, stronger in influence than the rift-type movement and resulted in the dramatic closure of the Japetus some 360-370 m.y. ago, forming the Arctic-North Caledonian fold belt (Dott, Batten, 1976). There, uplifting later set in (or continued) indicating that some form of vertical convection was (still) taking place. A number of orogenic events in the Appalachian-Mauretanides-Variscan geosynclinal belt followed the Caledonian deformation phase, so that in late Carboniferous time the North-American European craton was welded together along these fold belts forming a single continental mass (Ziegler, Louverens, 1979). A nearly continuous uplift of the new "old Red" continent, spreading its sediments into a large area, indicates that convective processes at depths were not totally dead. Moreover, as mentioned before, below all orogenic areas created in a compressional regime, the continental lower crust consists of : 1) much sialic material with a high degree of radioactive heat production ; 2) a low melting temperature ; and 3) a large thickness. These three factors must result in an accumulation of heat in the crust and even in the asthenosphere. As a consequence, a number of plumes will be formed in such low viscosity areas. Especially, the asthenosphere and the lithosphere below large orogenic belts must be special zones of weakness.

Certainly, the intensification or a new onset of a rift-type convection pattern is one of the most problematic topics in geodynamics, in contrast to the steady state of sea floor spreading and plate movement which is fairly well understood and explained (Turcotte, Oxburgh, 1967 ; 1976 ; McKenzie, 1978 ; Elsasser *et al.*, 1979). In general, in the beginning of a rift-type pattern more mass from below must be lifted upwards than is laterally carried away, producing an uplift as the first stage of rifting.

Regarding the physical details, so many uncertain quantities exist regarding depths, diameter, velocity, and temperature of a rising plume, that a very simplified mathematical treatment seems to be adequate and will be considered in the following section.

ASSESSMENT OF PHYSICAL PARAMETERS FOR THE PREPARATORY STAGE OF RIFTING

Before the period of stretching can set in, it is assumed that a number of loosely connected plumes of say 30-100 km in diameter with excess temperatures of typically 100-200° rise from various depths through the asthenosphere. An intensification or a new beginning of upward convection during the preparatory stage may involve areas within a rising plume, where not an adiabatic but a more iso-volumetric rise in pressure with temperature takes place. In these areas the increase in pressure before or at the beginning of uplift, rifting, and horizontal movements may be characterized by the extreme value :

$$\frac{\Delta P}{\Delta T} = \gamma \cdot \rho \cdot C_v = \alpha \cdot K_T \, ; \qquad (1)$$

p = pressure, T = abs. temperature ;
γ = Grüneisen-Parameter ;
ρ = density ;
C_V = specific heat at constant volume ;
α = coeff. of volume expansion ;
K_T = bulk modulus.

(see, for instance, Stacey, 1977).

Setting $\alpha = 3.10^{-5}(°K)^{-1}$; $\Delta T = 100°$, and $K_T = 10^{12}$ dyn/cm^2 one arrives at an excess pressure of 3 Kb. When released, this would be enough to initiate a total uplift of up to 10 km. Alternatively, one may estimate the buoyancy, velocity, and stress of a rising plume by using the equation of adiabatic volume expansion :

$$V_2 = V_1 \, (1 + \Delta T) \qquad (2)$$

or

$$\frac{V_2}{V_1} = \frac{\rho_1}{\rho_2} = (1 + \alpha\Delta T) \, . \qquad (3)$$

Solving for $\Delta\rho = \rho_2 - \rho_1$, setting $\rho_1 = 3.3$ gr cm^{-3} one arrives at $\Delta\rho = 0.01$ g/cm^3 for $\Delta T = 100°$. Using Stoke's formula of a rising sphere,

$$\eta W_o = (2/9) \; g \; R^2 \Delta\rho \qquad (4)$$

and calculating the shear stress,

$$\sigma = \eta \cdot \dot{\varepsilon} = \eta \cdot w_o/R = \frac{2}{9} g R \, \Delta\rho \qquad (5)$$

with g = acceleration of gravity ;
R = radius of sphere ;
η = viscosity of surrounding medium ;
ε = strain rate = w_o/R ;
w_o = average vertical velocity ;

one arrives at a value of σ = 7 to 20 bars for R = 30 to 100 km diameter for $\Delta T = 100°$ and σ = 13 to 42 bars for $\Delta T = 200°$.

As mentioned before, a pure adiabatic rise certainly does not take place in the *beginning* of plume formation, hence the stress values calculated are certainly a minimum. Similarly small figures are derived by Stacey (1977) on the basis of calculating the convective power of steady state mantle convection.

It is strongly supposed that nature will find a compromise between the two extremes of isovolumetric (equation 1) and adiabatic (equation 2) conditions in a rising plume. Hence, stresses of some hundred bars seem reasonable. Calculations have shown that their impact on the gravity field will be small for depths larger than 50 km and hardly detectable. This includes even the two extreme conditions.

In the following 4 sections, models of the development of a linearly elongated plume are demonstrated in 4 cross sections. The elongation might be a direct consequence of the linear zones of weakness of former orogenies.

Possible external forces which may influence the different stages of the development will be discussed in a separate section of this paper.

THE DEVELOPMENT OF STAGE 1, "UPLIFT" (Fig. 2.1 and 3.1)

A plume having crossed the asthenosphere and being under a certain overpressure in the order of some 100 bars is faced with an increasing viscosity toward the top of the mantle

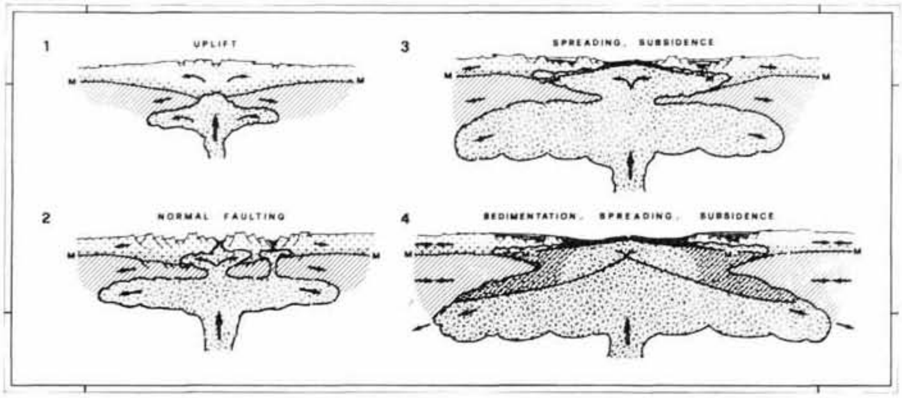

Figure 2

Model of a developing convection pattern : 2.1 Uplift stage : plume spreads into the asthenosphere and approaches the lower crust. Initiation of extension. 2.2 Extensional stage : stretching and block faulting in the lithosphere is caused by lateral movement of plume material into the asthenosphere ; plume material also spreads into the lower crust and leads to magma breakthrough. 2.3 Sinking and spreading stage : phase lag between intruding plume and lithosphere comes to a halt. Also the shear stress disappears. Spreading moves the separated continental parts aside (successful rift development). 2.4 Steady state spreading stage : sinking and sedimentation continuous at the newly formed ocean floor, particulary at the margins. New Moho is formed at shallower depth below the margins. Compression is built up in the continent outside the margins.

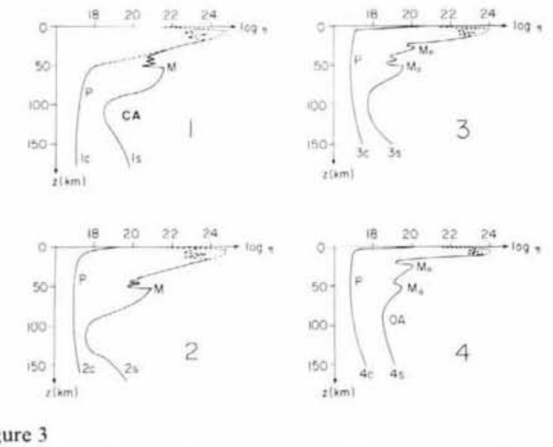

Figure 3

Development of the viscosity regime in the area of a rising plume : 3.1 Viscosities in rising plume (P) and in its neighborhood, stage 2.1 ; C = center of plume ; s = side ; M = Moho ; CA = continental asthenosphere. 3.2 Viscosities according to stage 2.2 : continental asthenosphere changes into oceanic asthenosphere. 3.3 Viscosities according to stage 2.3 ; magma breakthrough at center, beginning of formation of new Moho (Mn : new Moho ; Ma : old Moho). 3.4 Viscosities according to stage 2.4 ; slow cooling at asthenospheric levels below margins (OA : oceanic asthenosphere).

while there is not much resistance laterally, compare Figure 1. The lateral intrusion of material into the continental asthenosphere and a regional — not a local — uplift is seen at many juvenile continental rifts, e.g. in the surrounding of the Upper Rhinegraben, where at least 200 km on each side show an elevated Moho below the uplifted areas (Giese, 1976 ; Meissner, Vetter, 1974). It may be that a weak horizontal part of a convection cell was already established, or it may be that the uplift of the lithosphere on top of the plume strongly enhances the possibility of a lateral flow of plume material ; in any case the lateral movement will be much faster than the vertical movement, at least at this stage of plume development when the lithosphere is hardly disturbed. Regardless of the amount of overpressure below the dynamically elevated lithosphere, the shear stress of the vertically moving branch σ_{xz}, must be equal to or at least of the same order of magnitude as the shear stress of the horizontally intruding plume material, σ_{zx}.

From this relation one may formulate

$$\sigma_{xz} = \eta_L \cdot w_L / R = \sigma_{zx} = \eta_A \cdot u_A / \Delta h_A, \tag{6}$$

the indices L and A standing for lithosphere and asthenosphere.

u = horizontal velocity ;
w = vertical velocity ;
h_A = thickness of asthenosphere.

It follows from (6) that

$$\frac{u_A}{w_L} = \frac{\Delta h_A}{R} \cdot \frac{\eta_L}{\eta_A}. \tag{7}$$

As $\Delta h_A/R$ is of the order of 2 to 5 and $\eta L/\eta_A$ = 10 to 100 as shown in Figures 1 and 3.1, it follows that u_A/w_L is between 20 and 500. The stronger the viscosity gradient from the uppermost mantle to the asthenosphere, the larger will be the ratio of the horizontal to the vertical velocities of the deformed plume and the smaller will be the initial uplift. This might apply to purely oceanic rifts, which, however, will not be the main subject of this paper. Because of the lower viscosity in the asthenosphere, the overwhelming part of plume material moves sidewards introducing a strong shear stress at the upper part of the asthenosphere into the lithosphere. It is postulated that this stress is the reason for the "initial period of stretching". With a further increase of convective movement there will be a sub-lithospheric erosion and an intensified stretching by means of edge dislocation and formations of listric faults. The hampered upward movement of the plume material may now be able to reach the crust.

THE DEVELOPMENT OF STAGE 2, "NORMAL FAULTING" (Fig. 2.2 and 3.2)

At this stage the stretching and block faulting, mentioned in the last paragraph, assumes its maximum value because of the intensification of lateral movements in the asthenosphere. The plume material, entering also the lower part of the continental crust of a former orogeny (e.g. the Caledonian), gets access to an area of another — though smaller — viscosity minimum. As mentioned in the introduction, this viscosity minimum in the lower crust is especially strong in areas of thick sialic material. Plume material, still supposed to be under a slight overpressure, takes advantage of the lower crust's ductility and intrudes it laterally in a similar way as it intruded the asthenosphere before. It helps break the rigid upper crust and enters it along one or several prominent fault zones, while the stretching, caused by the establishment of strong horizontal asthenospheric movements, provides the space in the lower crust where plume material replaces more and more the former continental material. Such a migration of mantle material to the lower part of the crust of continental margins is also postulated by Kosminskaya and Pawlenkova (1979), on the basis of many seismic sections at various margins of Eurasia. New evidence by Montadert *et al.* (1979) and Le Pichon and Sibuet (in press) shows the strong tendency that block faulting in the upper crust is listric and tends to assume a more and more horizontal direction in the (ductile) lower crust. At the end of stage 2, characterized by extreme crustal extension basaltic magma, may not only enter the lower crust but also form huge extrusions, overflowing adjacent continental crust, as found on the Vøring Plateau (Eldholm, Sundvor, 1979).

THE DEVELOPMENT OF STAGE 3, "SINKING AND SPREADING" (Fig. 2.3 and 3.3)

The phase lag between the intruding and horizontally spreading plume material and the so far static lithosphere decreases when the former continental unit finally is split and moves laterally to both sides of the ridge. Now, a nearly continent-wide convection has formed, and the split continental units ride on the underlying asthenosphere, there by coming in phase with the horizontal movement below. This means that the stress σ_{zx} approaches zero, stretching and block faulting stop, and the stage of steady state sea-floor spreading is reached. It also means that no more plume material enters the lower continental crust, and cooling processes start to dominate. In the area of the former horizontal stresses, cooling and contractions prevail and lead to subsidence within the whole area which was intruded by plume material before. This includes the shelf areas. Instead of the former extension and block faulting now a quite subsidence process with an incipient sedimentation is initiated. A new higher moho is formed by gravitational differentiation.

Many linear and young spreading centers were formed during the Early Cretaceous in the Mesozoic Arctic-North Atlantic part of Pangaea (Ziegler, 1978). Only those below former orogenic belts survived, the others getting into the stages of "failed rifts". Whether a mutual interaction of the young rifts or an overlying convection/stress system — possibly caused by processes at active margins around Pangaea — was involved in the forming of only one or two spreading rifts remains doubtful and will be discussed again in a later section.

THE DEVELOPMENT OF STAGE 4, "SPREADING, SINKING, SEDIMENTATION" (Fig. 2.4 and 3.4)

Subsidence and sedimentation below the shelf areas continue due to cooling and contraction of the lithosphere, because of the increasing sediment load, and of possible phase changes from gabbro to eclogite. The new Moho below the shelf may join the adjacent continental Moho, but in a transition zone two separate Mohos may be observed. The sea floor spreading will find more and more lateral resistance in the adjacent continental areas where compressional stresses might become rather strong. Such stresses have been observed in Norway and Sweden by Hast (1977). The regional intrusion of low-velocity plume material at asthenospheric levels is reflected by large travel time residuals, e.g. those of Scandinavia which have delays of up to 1 sec. between the Norwegian coast and Central Sweden (Weinrebe, 1981), and by the Caledonian gravity low in an area without a crustal root (Theilen, Meissner, 1979). Tendencies of the oceanic asthenosphere to further intrude the continental asthenosphere may show more and more vertical components of movement, a precursor of a possible future subduction.

THE VISCOSITY MODELS (Fig. 3.1 to 3.4)

In accordance with the models of Vetter and Meissner (1977 ; 1979) based on selected geotherms and Weertman's temperature method (Weertman, 1970 ; Kohlstedt, Goetze, 1974 ; and others), the viscosity structure before the development of a plume was already introduced in Figure 1. The following Figures 3.1 to 3.4, each showing the viscosity structure in the center of the rising plume and in its neighbourhood, approximately 30 tp 50 km from the center, give the different stages of vertical and horizontal changes of the viscosity structure in accordance with Figure 2.1 to 2.4. It is stressed that the curves were plotted on the basis of simple temperature estimates.

Modelling by finite element calculations has begun but is still in a preparatory stage.

DISCUSSION

While some important indications for the validity of the proposed development were already noted together with the description of the models, some more arguments will be discussed. The models are based on the viscosity structure which strongly determines the convection paths.

The old questions regarding the depth of convection patterns should not be broadly discussed again. Only 3 observations will briefly be mentioned : certainly the depth of earthquakes and displacements in subduction zones down to 700 km, the forming of a new ocean with huge masses of new lithosphere, and the tremendous forces necessary for pushing large continents away strongly argue for a deep seated convection. Moreover, viscosity estimates (Weertman, 1970 ; Vetter, Meissner, 1977 ; Meissner, Vetter, 1976) indicate the possibility of creep convection in the whole mantle. The question, however, remains, whether the force for pushing the continents away comes predominantly from the upward push of the convection branch or what other forces have to be considered. Could it be a world-wide overall stress pattern which causes the stretching of the lithosphere in the first place ? Subduction zones with its pull are far away in the middle of Pangaea. A stress in the lithosphere would have to be relayed along a rigid plate of a thickness to length ratio of 1/100. This seems impossible without breaking the plate at many places. In the asthenosphere stress is connected with strain rate, i.e. with creep and convection again. Whether or not such an overall convection contributes to the first period of stretching and forming of rifts might be estimated by plotting the intensity of rifting as a product of direction and length from Ziegler's rift pattern map (Ziegler, 1978) of the North Atlantic area at the beginning of the opening in Early Cretaceons time (Fig. 4). No strong preference of direction is observed. The slightly dominant NS direction agrees with the strike of the Caledonian orogeny, as mentioned before. Rifts developed however in a large area and in mostly different directions. This observation alone strongly argues for a deep-seated process. The final alignment in a preferred NS direction must be attributed to the low-viscosity area below the Caledonian orogenic system and a possible contribution of a former convection system in this area, as mentioned before. The observation that ocean ridge basalts are generally depleted in the Strontium ratios 75 Sr/86 Sr, (Engel *et al.*, 1965) means that they do not come from a juvenile untouched mantle, but have suffered (repeated) differentiation processes in the past.

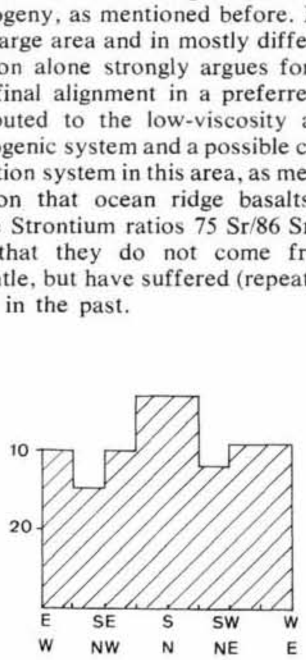

Figure 4
Number of rifts x length for the Mesozoic North-Atlantic rift system (after the map of Ziegler, 1978).

The presented model of rift and margin formation is in agreement also with many geological and geophysical data from specific areas. Figure 5 is an example from a new refraction-reflection line across the continental margin of Northern Norway (Avedik *et al.*, 1981). The faults of the basement, apparently listric at depth, and the undisturbed sediments representing the different stages of the model are

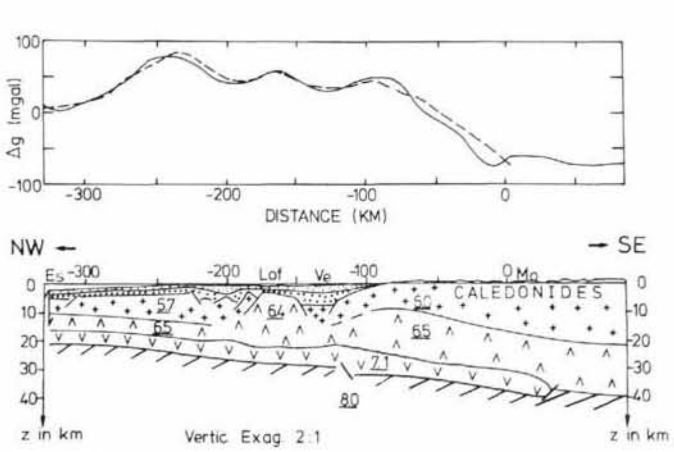

Figure 5
Cross section through Norwegian Margin with geologic units, seismic velocities and gravity model : ——— = Free air anomaly, as measured by Meteor ; — — — = theoretical gravity ; Es = Vøring Plateau Escarpment ; Lof = Lofoten ridge ; Ve = Vestfjord basin ; Mo = Mo I Rana.

clearly defined. The Vestfjord basin appears as one of the many "failed rifts" of the North Atlantic while the old block of the Lofoten remained undisturbed during the period of stretching. Thinning and stretching apparently has also influenced the velocity of the basement which is much lower below the shelf than on the "real" continent. It is supposed that micro- and macro-cracking of the basement has reduced the velocity of seismic waves as known from laboratory measurements. As a very interesting feature a layer of 7 to 7.2 km/s has been found at the base of the crust. It can be interpreted that the residue of the plume material which intruded the continental crust during the stretching period, formed the new Moho which is observed to continue from here to the ocean basin.
As discussed above on the landward side a long-wavelength gravity low and teleseismic delays are observed. Their origins certainly are at asthenospheric levels. Farther to the east compressional structures are found, originating — according to the model — from compressional stresses in the end region of the intruding plume. Here also the teleseismic delays disappear. Presently, other continental passive margins are tested with regard to the validity of the model.

CONCLUSIONS

The model as presented in Figures 2.1 to 2.4 gives a general idea on the formation of rifts and passive margins. It is compatible with the majority of geological and geophysical observations. The general regional uplift in the preparatory stage and the following period of stretching and block faulting are interpreted directly as a consequence of rising plumes in a medium with a viscosity gradient in the upper mantle. The period of subsidence and quiet sedimentation starts when the continent is finally split and the phase lag between the laterally intruding plume material and the overlying lithosphere comes to a halt. Plume material which has entered the lower continental crust forms a new crust-mantle boundary by gravitational differentiation. The new Moho is much higher than the older one and presents a natural continuation of that formed by the rift process. On land, the intruded plume material in the asthenosphere causes large travel time delays and negative gravity anomalies with respect to the unintruded continent. In a transition zone below the continent compressive stresses are built up. Generally, most observations around passive margins are understood as a heritage of the rift process which had its main activity at depths. Cooling with contraction, sedimentation, and possible phase transitions are rather well described processes explaining the mature stage in the development of the passive margins. Such margins may at future stages initiate a downward movement and a decoupling of the oceanic plate. Together with a downward-directed component of the asthenospheric flow a passive margin may then turn into an active one. This development is part of Stille's description of orogenies (Stille, 1944 ; Kraus, 1959). The physical explanation of these features of global tectonics lies in the fact that the earth is a great heat engine. Rising material and the push from the hot and elevated ridges play an equally important role as do the sinking and the pull at the cold zones of subduction. The viscosity structure and its relation to temperature only provide the boundary conditions for the sometimes complicated pattern of convection.

Acknowledgements

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