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Viscosity-Depth-Structure of Different Tectonic Units and Possible Consequences for the Upper Part of Converging Plates*

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Abstract. The ratio of melting point to temperature is considered as a measure for the effective viscosity. From deep seismic sounding and high pressure — high temperature studies certain conclusions on the material of crust and upper mantle are derived. They are the basis for estimating melting point curves. From these data and from temperature-depth curves viscosity-depth functions are obtained. The difference in viscosity between adjacent oceanic and continental structures when under tension or compression seems to be responsible for a number of observed features such as gravity highs and mafic outcrops behind trench regions.

Key words: Viscosity-Depth-Structure — Crust — Continental Margins.

1. Introduction

The determination or estimation of viscosity and rigidity of the outer parts of the earth is a difficult task and has been treated in various ways. Absolute values of viscosity on a regional scale have been obtained from the uplift of the Fennoscandian Shield (Haskell, 1935), the Canadian Shield (Berry and Fuchs, 1973), also from estimations of the quality factor Q based on absorption measurements of seismic body waves (Anderson, 1967). These and other approaches mostly lead to viscosity values as an integral part of large areas and depths intervals, although from tectonic and seismic evidence it can be concluded that the viscosity in the low velocity zone (or astenosphere) is smaller by a factor of 10 to 100 than in the layers above and below (Lliboutry, 1972).

The first attempt to obtain a viscosity depth-function for the mantle was made by Weertman (1970). In his excellent review paper on the creep strength of the mantle all possible uncertainties are mentioned which are involved in the derivation of rather reliable viscosity values. It seems

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strange that so far no attempt for estimating crustal viscosity-depth profiles has been made, especially as our knowledge on crustal material, on its temperature and melting points is at least better than that on the mantle.

However difficult the development of reliable viscosity data may be there is a general agreement that viscosity and creep play a major role in all tectonic processes. A fact of great importance for geodynamics is that brittle materials such as nearly all rocks tend to become ductile when subjected to compression from all sides. It appears that for confining pressures up to about 4 kbar (corresponding to depths of about 12 km) most crustal rocks behave as a brittle material, but for confining pressures of 6 kbar and over (depths of 20 km) they are capable of considerable plastic deformation. Moreover, long time steady state creep may occur at all stresses, although its rate increases greatly with increasing stress and with increasing temperature. The creep rate $\dot{e}$ at constant stress, according to Jaeger (1969), may be expressed by a simple relation:

$$\dot{e} \sim \exp(-A/k \cdot T) \qquad (1)$$

A = constant value depending on chemistry and mineralogy of a rock
k = Boltzmann constant
T = absolute temperature

In a perfectly viscous fluid or "Newtonian" substance thre creep rate $\dot{e}$ may be expressed as

$$\dot{e} = \tau/\eta \qquad (2)$$

τ = stress
η = viscosity

Hence, $\dot{e}$ is reciprocal to the viscosity which is true not only for a Newtonian body but for other models such as Kelvins' or Binghams' in the long time approach as well. In natural rocks near the surface viscosity values may vary from 10^{22} poise (gabbro, 20° C) to 10^4 poise (lava flow, 1100°), so covering a wide range of viscosity values even for the same material but, of course, at different phases and temperatures ranges.

From Weertman's thorough investigations it seems probable that at stresses greater than 10^{-2} bar dislocation creep is the dominant creep process. Glide controlling mechanisms lead to refined creep equations such as

$$\dot{e} = \alpha_1 \cdot D(\tau/\mu)^2(\tau\Omega/kT) \text{ with} \qquad (3)$$

$$D = D_0 \exp(-\alpha_2 \cdot \Theta/T) = \text{Diffusion coefficient} \qquad (4)$$

α_1 = constant 1	τ = stress
α_2 = constant 2, about 18	μ = shear module
Θ = melting point	Ω = atomic volume
T = temperatur	k = Boltzman's constant

Rewriting (2) as

$$\eta = \tau/\dot{e} \tag{5}$$

with $\dot{e}$ from (3) and D from (4) it is evident that the diffusion coefficient D, the creep rate $\dot{e}$, and the viscosity η strongly depend on the ratio Θ/T. For instance, we have $D = 10^{-8}$ cm^2 sec^{-1} at $T = \Theta$ and $D = 10^{-16}$ cm^{2}sec^{-1} for $T = \Theta/2$, according to Weertman (1970) who also introduced the "effective viscosity" $\tilde{\eta}$ as the viscosity at a constant creep rate of $\dot{e} = 10^{-16}$ sec^{-1}. This "effective viscosity" seems to be a very efficient tool for comparing viscosity values in crust and upper mantle. The relation of $\tilde{\eta}$ to Θ/T is so strong that it is even possible to write approximately:

$$\tilde{\eta} \approx \alpha_3 \exp(\alpha_2 \, \Theta/T) \quad \text{or:} \quad \ln \tilde{\eta} \approx \ln \alpha_3 + \alpha_2 \, \Theta/T, \tag{6}$$

From this simple relationship it is clear that one has to derive values of Θ and T of representative crustal and upper mantle units in order to estimate the proper viscosity values. Both Θ and T strongly depend on the material in crust and upper mantle. Other relationships and uncertainties of Θ and T will be discussed in a later chapter. But, as it is strongly suggested that $\Theta(z)$ and $T(z)$ are indeed very different in various tectonic units, as for instance in shelfs, shields, oceans, and young montain systems, it seems challenging to investigate the viscosity behaviour of these various areas and their viscosity differences in more detail. Because of the strong relationships between seismic velocity and material and between material and melting point in a given pressure-temperature environment first the crustal material compatible with deep seismic sounding data will be discussed.

2. Seismic Data of Crustal Structure, Connection between Seismic Velocity and Material

The relation between seismic velocities and the material (or composition of rocks) has been treated by a large number of authors. (See Birch, 1960, 1961; Christensen, 1965; Hughes and Maurette, 1956; Meissner, 1967). Combining these data with velocity-depth structures as obtained from detailed seismic investigations it appears that the upper part of continental crusts down to about 10—12 km consists of granitic (or sialic) material with seismic velocities around 5.7—6.3 km/s. Whereas crustal low velocity zones in the Alps and other young mountain belts indicate that sialic material may reach depths of 20 or more kilometres, shield areas may consist of more gabbroic, possibly dioritic material at medium crustal depth (Vetter and Meissner, 1970). Up to now, the deepest part of continental crusts in general was considered to consist mainly of gabbro, however according to investigations of Ringwood (1969) and Ito and Kennedy (1971), a phase transition to garnet granulite (and hornblende) should take place at p—T conditions equivalent to those of lower continental crusts. The transition from crust to upper mantle seems to be stepwise in an overall velocity gradient zone. Velocities larger than 7.8 km/s indicate that the Mohorovičić-Discontinuity (= Moho) has been reached. For a number of reasons (Meissner, 1973) peridotitic material should prevail in the uppermost part of the mantle where in general average velocities of 8.0 to 8.2 km/s are found. This dense peridotitic or pyroxenitic material has a high melting point and a high viscosity.

Compared to continental crusts the oceanic crust is much more simple and shallower. In the average, below 5 km of ocean water it is only 5 km thick reaching from 5 to 10 km. Underneath thin layers of sediment and basalt there has gabbro to be assumed for the bulk of the crust. The uppermost part of the mantle, beginning already at 10 km depth, seems to consist of peridotitic material, similar to the mantle material of continental crusts

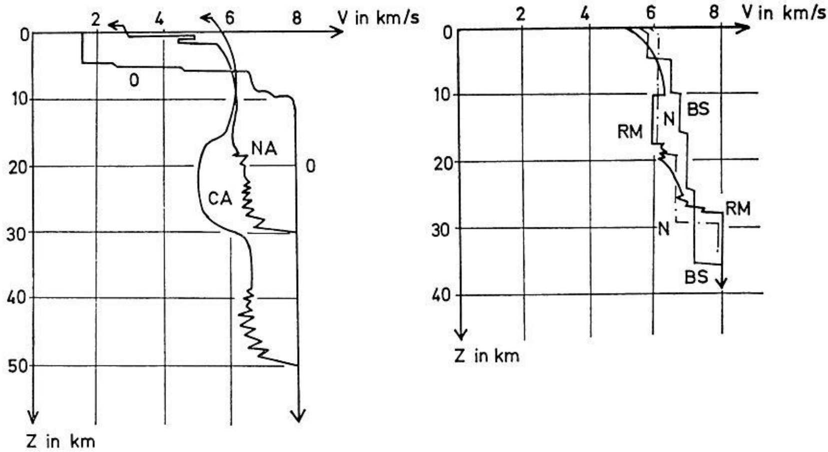

Fig. 1. Velocity-Depth Curves. *left:* Typical V—z functions obtained with modern evaluation methods. *right:* V—z functions from shields and old mountain systems, continental areas

O = Ocean Floor
NA = Northern Boundary of Alps and Molasse
CA = Central Alps
≷ = Zone of lamellation from seismic data

BS = Shield areas: Baltic Shield
N = Caledonian system: Norway
RM = Hercynian system: Rhenish Massif

with their Mohos at deeper and most different depths. It is already known that the oceanic crust and upper mantle are much colder and hence much more rigid than their continental counterparts. Some velocity-depth profiles of representative areas of young mountain belts (Central Alps), from the boundary of the Alps (Northern Alps), and from an average oceanic crust are shown on the left part of Fig. 1. Some more data of shield and shelf areas are presented at the right part of that figure. It can be seen that the same depth ranges inhibit a completely different velocity structure. Differences are especially large between oceanic and young mountain areas. As will be shown these differences are not only found for the velocity structure and the material; also the viscosity turns out to be very different at comparable depth ranges which may play an important role when adjacent units come under tension or pressure.

3. Assumptions for Temperature-Depth Functions and Melting Points

Temperature — depth functions (or geotherms) $T(z)$ are normally calculated on the basis of heat flow values and crustal structure. Reduced values are 0.92 ± 0.17 μcal $cm^{-2}sec^{-1}$ for average shield areas (Lee and Uyeda, 1965), 1.39 ± 0.4 for the Alps (Haenel and Zoth, 1973), and 1.28 ± 0.53 for average ocean basins. Using these average heat flow values still additional assumptions on the heat production and the thermal conductivity have to be made in order to derive $T(z)$ for certain regions (Buntebarth, 1973). So, temperature-depth curves should be taken with some caution although it is quite evident that temperatures in the crust and upper mantle follow a decreasing scale from volcanic areas, orogenic areas, ocean basins, to shield areas. Geotherms, of course, are smooth curves and even monotonically increasing for crustal depths.

Of similar uncertainties as the geotherms are the melting point curves $\Theta(z)$. They are a function of material, pressure, and saturation with liquids. The material of the upper crust and the upper mantle seems to be known fairly well on the basis of deep seismic sounding; the p—T relationship of melting points of different materials under various saturation conditions has been investigated in the laboratory by a large number of authors (Wyllie and Tuttle, 1960; Yoder and Tilley, 1962; Kushiro, 1968; Ringwood, 1969; Green and Ringwood, 1970; Wyllie, 1971; Ito and Kennedy 1971).

Fig. 2 shows the geotherms of characteristic areas together with melting zones of some natural rocks with water included. The melting point curve of granitic rocks crosses the geotherm of the Alps at about 20 km depth, that of gabbroic rocks between 30 and 40 km. These depths are still inside the crust of the Central Alps (= CA), i. e. above the Moho, indicated as M on

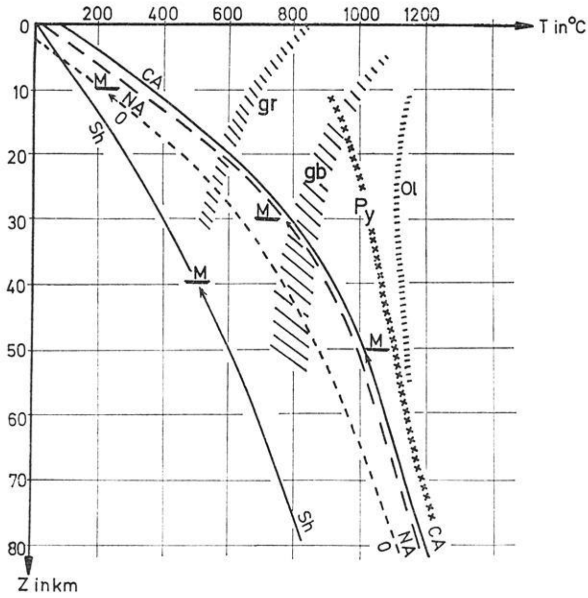

Fig. 2. Geotherms $T(z)$ and Melting Points $\Theta(z)$

Sh = $T(z)$ function of Shield areas (Ringwood, 1969)
O = $T(z)$ function of Oceanic areas (Ringwood, 1969)
NA = $T(z)$ function of the Northern Alps (Buntebarth, 1973)
CA = $T(z)$ function of the Central Alps (Buntebarth, 1973)
M = Mohorovičić Discontinuity
gr = $\Theta(z)$ of granitic rocks
gb = $\Theta(z)$ of gabbroic rocks
Py = $\Theta(z)$ of pyroxenitic rocks
Ol = $\Theta(z)$ of olivin
(after Wyllie and Tuttle (1960) and Yoder and Tilley (1956)

the geotherms. For the other regions an intersection between melting point- and geotherm will not take place in reality as granitic or gabbroic material most probably do not reach these depths. Figs. 3 and 4 show more recent investigations on the gabbro-eclogite and on the peridotite transitions.

In addition to the geotherms from Fig. 2, the wet solidus curve (= WS) and the dry solidus curve (= DS) are shown. From these data of Ito and Kennedy (1971) (Fig. 3) no gabbroic rock should exist in the deeper continental crust. The geotherms CA and NA both cross a large zone where only garnet granulite and hornblende can exist, materials, which have turned out to be stable in this p—T region. A transition to eclogite could only happen for very deep Mohos, as for instance for the CA-curve (see Fig. 3).

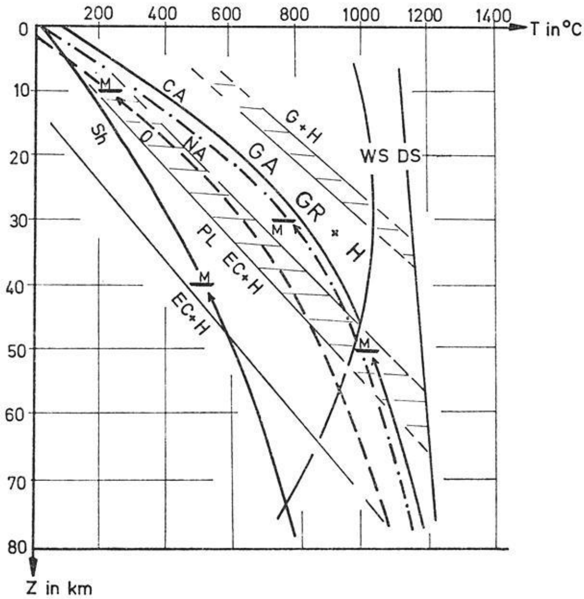

Fig. 3. Geotherms $T(z)$ and Phase Relationship for Gabbro (Ito and Kennedy, 1971; Wyllie, 1971)

Sh, O, NA, CA, M } see Fig. 2
WS = Wet Solidus

DS = Dry Solidus
G + H = Gabbro + Hornblende
GA GR + H = Garnet Granulite + Hornblende
PL EC + H = Plagioclase Eclogite + Hornblende
EC + H = Eclogite + Hornblende

According to this diagram all other geotherms are not compatible with an eclogitic upper mantle. Gabbroic rock, from these investigations, can only occur at shallow depth where the reaction velocity between the solid phases is very slow. The transition zones between the solid phases according to Ito and Kennedy (1971) are smaller than those of Ringwood (1969). In contrast to these rather strong phase and density changes of gabbroic material those of peridotitic rock (Fig. 4) are comparably smooth and do not show much changes. So, peridotitic material is compatible with all geotherms. An intersection between geotherm and melting curve does not occur before depths of more than 80 km are reached. As melting point-depth curves strongly depend on the material they do in general not increase monotonically.

In order to come to an estimation on the effective viscosity the ratio between the melting point curve $\Theta(z)$ and the geotherms $T(z)$ has been

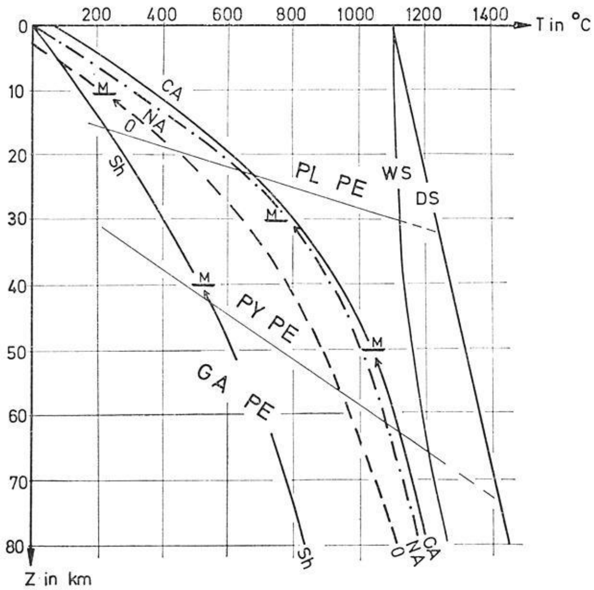

Fig. 4. Geotherms $T(z)$ and Phase Relationship for Peridotite (Green and Ringwood, 1970, Kushiro *et al.*, 1968)

Sh, O, NA, CA, M } see Fig. 2

WS = Wet Solidus
DS = Dry Solidus

PL PE = Plagioclase Peridotite
PY PE = Pyroxene Peridotite
GA PE = Garnet Peridotite

plotted in Figs. 5 and 6. Always the wet solidus curves of Fig. 2, 3 and 4 have been taken for $\Theta(z)$ in order to have the same reference curve. Wet granitic material is assumed for the upper part of continental crusts, gabbroic material for the whole oceanic crust. In the deeper part of continental crusts two possibilities with regard to the material have been taken into account, and for the upper mantle pyroxenitic peridotite has been assumed.

In accordance with formulas (5), (4) and (3) also values of the effective viscosity have been calculated for a constant creep rate of $\dot{e} = 10^{-16}\mathrm{sec}^{-1}$. These obtained values of $\tilde{\eta}$ are labelled in the diagrams of Figs. 5 and 6 together with Θ/T. The stress which is necessary to obtain a constant creep rate will, of course, be much larger in zones of higher viscosity (and higher Θ/T values) than in zones of viscosity minima. Because always the wet solidus for Θ was used as a reference curve the indicated values of $\tilde{\eta}$ should be considered as the lower limit of viscosity values.

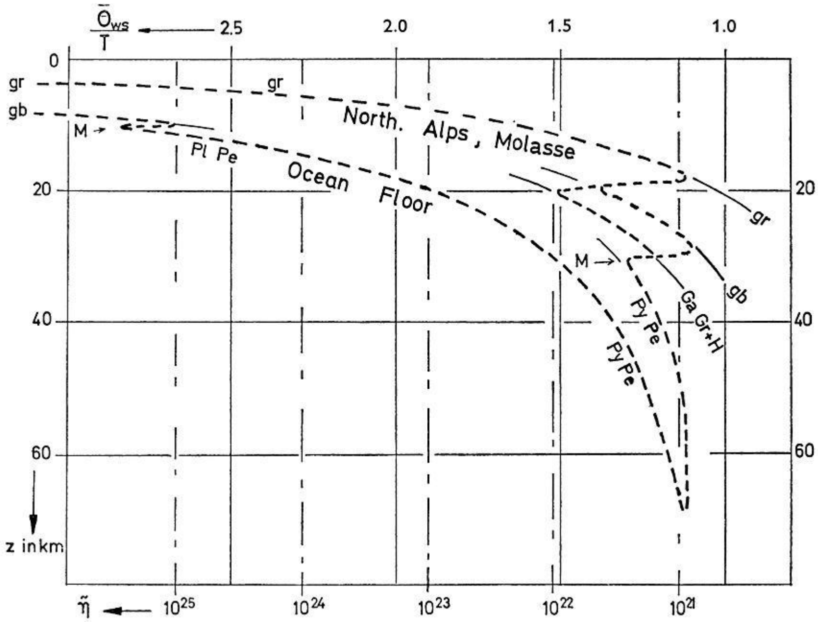

Fig. 5. Ratio Average Melting Point Θ_{ws} to Temperature T and Effective Viscosity $\tilde{\eta}$ as Function of Depth; Northern Alps + Molasse Basin and Average Ocean Floor; $\tilde{\eta}$ in poise

gr = granitic rocks
gb = gabbroic rocks
M = Mohorovičić Discontinuity
Py Pe = Pyroxenite Peridotite
Ga Gr + H = Garnet Granulite + Hornblende

For the northern boundary of the Alps and most probably for the bulk of Western and Central Europe two minima of the Θ/T curves are found: They are at around 18 and around 29 km depths. See Fig. 5. Comparing these curves with velocity-depth curves of Fig. 1, the upper viscosity minimum agrees with a zone of rather constant or decreasing velocity in the upper and middle part of European crusts, the lower minimum coincides with the — probably laminated — transition zone between crust and mantle.

In contrast to continental areas the Θ/T curve for oceans shows only one very weak minimum at the base of the crust and remains at a large distance from 1, indicating a very high viscosity for crust and uppermost mantle. Shield areas, Fig. 6, with their low temperatures show a still higher viscosity and again two minima of the viscosity-depth curves.

Here the transition from sialic to gabbroic material is at rather shallow depth and seems to be different in various shield areas. For the Central Alps also two minima of Θ/T are found, the upper one here around 22 km depth

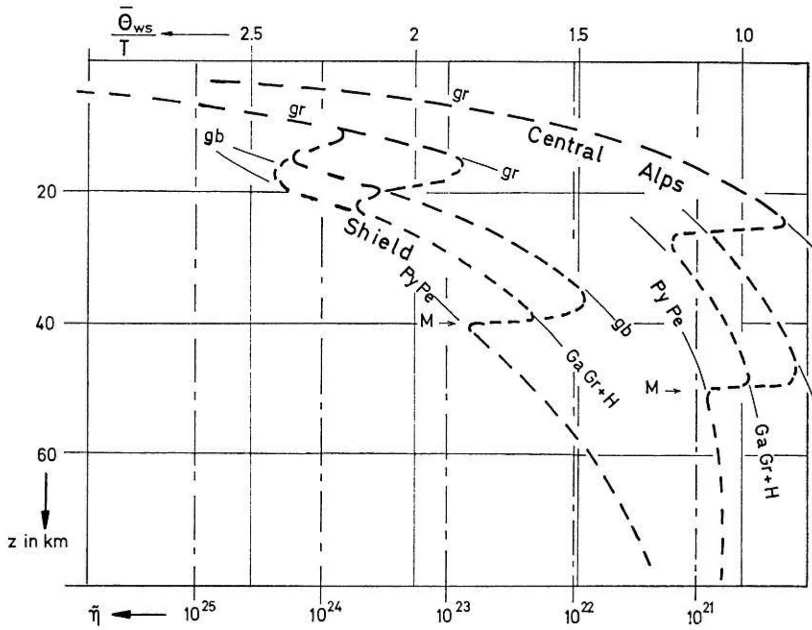

Fig. 6. Ratio Average Melting Point Θ_{ws} to Temperature T and Effective Viscosity $\tilde{\eta}$ as Function of Depth; Central Alps and Average Shield Areas

gr = granitic rocks
gb = gabbroic rocks
M = Mohorovičić Discontinuity
Py Pe = Pyroxenite Peridotite
Ga Gr + H = Garnet Granulite + Hornblende

with a possibility for melting if wet granitic material is present at these depths. If wet gabbro should exist in the lower part of the Central Alps an extensive melting zone would be present. But according to data from Ringwood (1969) and Ito and Kennedy (1971) this seems rather improbable. The upper viscosity minimum in the Alps agrees with a pronounced low velocity zone between 18 and 28 km, the lower zone with a rather large transition zone between crust and mantle. It should be stressed that these two minima appear at all continental crusts as long as there is a definite differentiation into a predominantly granitic, a more basic lower crust, and an ultrabasic mantle resulting in steps of the melting point curve and hence in the ratio Θ/T.

In Fig. 7 different symbols for $\tilde{\eta}$ have been plotted along the velocity-depth curves of Fig. 1, indicating different viscosity values for the four areas under consideration. Low velocity zones seem to coincide with low viscosity zones. In agreement with this concept extended and pronounced low velocity zones in the European crust so far have been detected with certainty only in the Alps and not in the adjacent shelf, shield, or oceanic regions.

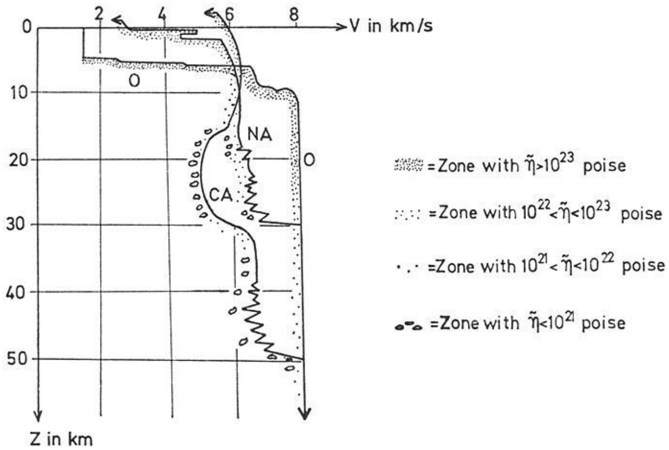

Fig. 7. Velocity-Depth Curves with Indicated Viscosity Values

O = Ocean Floor
NA = Northern boundary of Alps
CA = Central Alps

On the basis of gravity calculations and stability considerations it is quite probable that even in the Central Alps no large zones of complete melting will exist because the saturation with volatile liquid may not be complete. Certainly, however, the first and the second minimum of the Θ/T-curve will be present. If in some zones complete melting would take place the velocity of compressional waves in this zone would decrease to that of shear waves in the corresponding cool and solid material:

$$V_P = [(\lambda + 2\,\mu)/\varrho]^{\frac{1}{2}} \rightarrow (\lambda/\varrho)^{\frac{1}{2}} \text{ for } \mu \rightarrow 0 \text{ (V}_P \text{ in melts)} \tag{7}$$

As in solids: $\lambda^{so} \approx \mu^{so}$ and $V_S^{so} = (\mu/\varrho)^{\frac{1}{2}} \approx (\lambda/\varrho)^{\frac{1}{2}}$ it follows that:

$$V_P^{me} \approx V_S^{so} \approx (V_P^{so}/3^{\frac{1}{2}}) \text{ assuming } \lambda^{so} \approx \lambda^{me} \tag{8}$$

λ, μ = Lamès constants
ϱ = density
V_P^{me} = compress. velocity in melt
V_P^{so} = compress. velocity in solid
V_S^{so} = shear velocity in solid

From (8) we may conclude that velocities of granitic material with $V_P^{so} = 6$ km/s cannot have values lower than $V_P^{me} = 3.5$ km/s; velocities of gabbro or granulite with $V_P^{so} = 7$ km/s should not show velocities lower than $V_P^{me} = 4$ km/s for a melting zone. From seismic evidence in the Central Alps in some parts average velocities below 5 km/s have been found

(Giese, 1968; Meissner, 1968), so that limited zones — possibly laminated — of velocities around 4 km/s are quite probable. Hence, also viscosities in these zones should be extremely low.

4. Consequences for Lateral Movements at Adjacent Plate Boundaries (Continental and Oceanic Type)

From the viscosity-depth curves of Figs. 5 and 6 interesting clues may be obtained for areas where oceanic and continental units come together. Even without lateral compression adjacent units of crustal and uppermost mantle down to 80—100 km — now to be called: the upper part of plates — may show tendencies of lateral movements as pressure differences up to about 1 kbar may be present. Assuming two columns of upper oceanic and upper continental plate as represented by the curves O and NA in previous figures, a curve of pressure differences versus depth may be obtained using the following formula

$$\Delta p_{c-o} = g\left(\sum_{c} \varrho_i h_i - \sum_{o} \varrho_i h_i\right) \qquad (9)$$

ϱ_i, h_i = density and depth intervals

Δp_{c-o} = pressure difference between continental and oceanic plate

In Fig. 8 left two adjacent density columns and a Δp versus z — graph are presented.

The $\Delta p(z)$-curve shows a maximum at about 5 km depth and positive Δp values from zero down to about 30 km depth. At these depth intervals there should be a tendency for continental areas to move or creep against its oceanic neighbour, a process which may be obscured by a sedimentary movement from continent to ocean because of erosion. From about 30 km downward small negative Δp values are found as here the cooler and more viscous oceanic plate is denser than the continental one. It should be noted that adjacent continental and oceanic plates, not separated by large shelf areas, would show some gravity maximum at the continental and a minimum at the oceanic side of their boundary although both areas may be in isostatic balance.

It seems possible that pressure differences, as shown in Fig. 8 left, may initiate or at least support those lateral movements which are observed at subduction zones. Stacey (1969) supposes that steady state creep plays a major role for depths larger than 10—15 km. This depth range according to Figs. 5 and 8 right is characterized by the largest viscosity differences. If lateral compression is applied to this system faulting and folding should start at the more viscous parts, i.e. in the uppermost parts of crust and

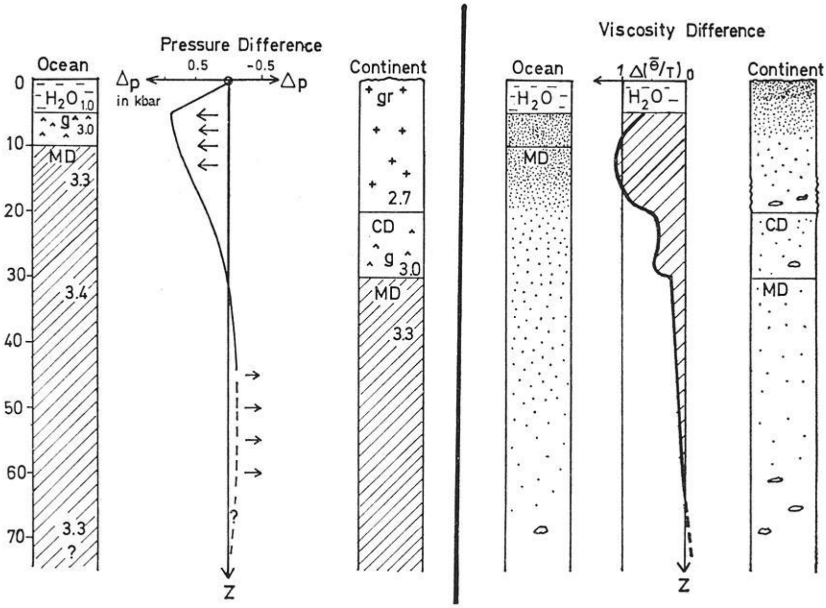

Fig. 8. Pressure Difference and Viscosity Difference of Adjacent Oceanic and Continental Areas; CD = Conrad-, MD = Mohorovičić-discontinuity

left: Density distribution $\varrho(z)$ and difference of hydrostatic pressure Δp between continental and oceanic areas

right: Viscosity structure and difference in viscosity $\Delta(\Theta/T)$ between oceanic and continental areas

mantle. Creep should occur in the low viscosity regions of the crust (and — of course — in the astenosphere), as indicated in Fig. 9, top. With the development of stronger compressional movements there should always be a tendency that pieces or lamellas from the cold, rigid, and viscous uppermost part of the oceanic plate intrude and get incorporated into the softer and less viscous parts of the continental lower crust. See Fig. 9, bottom. Especially parts of the viscous oceanic transition zone crust-mantle will stick together and will be transferred to the softer and increasingly thicker and warmer granitic continental crust. This means that definitely not the whole oceanic plate moves below the whole continental plate, especially, as the upper part of the rigid oceanic crust and mantle is at the same depth as the weak continental crust. Large intrusions which are dynamically pressed into the softer crusts of the developing orogeny must cause large gravity highs. This process may be the reason for the forming of the 5000 km long belts of strong positive gravity anomalies in the western part of the Andes, the gravity high of the Ivrea Body in the Alps, and other similar anomalies near plate boundaries. During the process of compression

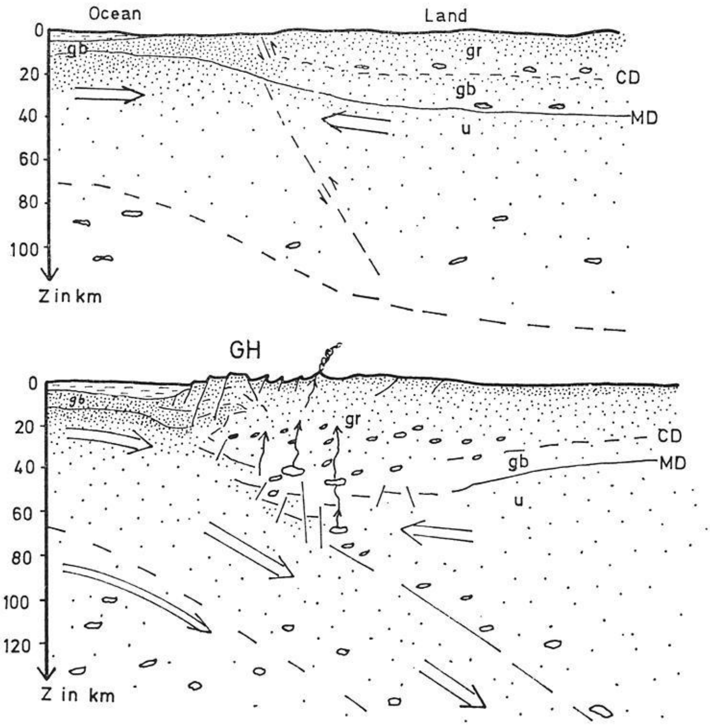

Fig. 9. Viscosity Structure at Continental Margins

upper part: Rather passive margin, begin of compression
lower part: Advanced stage of subduction
GH = Gravity High   gr = granitic layers
CD = Conrad Discontinuity   gb = gabbroic layers
MD = Mohorovičić Discont.   u = ultrabasic layers
Viscosity structure as in Fig. 7

a forming of horsts, overthrusting, and even the appearance of peridotitic (and gabbroic) spans from the mantle on the surface may take place while in the area of the trench sedimentation with normal faulting should predominate. As a further consequence of this concept, a faulting or peeling of the uppermost, viscous part of the (formerly) oceanic mantle should lead to an enhancement of lamellas or lenses connected with the Moho boundary. Also, several Mohos may develope by this process which may be stable for very long times.

Conclusions

Certainly, the ratio of melting point to temperature as a funtion of depth can not be determined uniquely. The largest uncertainty may be the rather unknown saturation of crustal rocks with volatile liquids. The generally decreasing water content with depth effects mainly the melting point curves so that for larger depth the dry solidus curve should be a better approximation to the true conditions. This would increase the Θ/T ratio and the viscosity values. Investigations of specific areas making some assumption regarding the decreasing saturation with liquids will be performed in the near future. In this respect the present study should only be considered as a first attempt to obtain the range of possible crustal viscosity values.

It should again be mentioned that steps in the melting point-depth curves will always happen at chemical boundaries inside the upper part of plates. These steps necessarily lead to steps and relative minima in viscosity and, hence, to a higher creep rate in these zones if stress is applied. Also the difference in viscosity between two adjacent geotectonic units will not be much different when using dry solidus instead of wet solidus curves. The geotectonic consequences will not be altered. When looking for zones of enhanced creep, deformation, and possibly for a decoupling between layers of different depth those zones of reduced viscosity (and velocity) have to be found.

It has been shown that the well known and much discussed features of plate tectonics at continental margins may be supplemented by a number of processes which are the result of a different physical status of adjacent but different upper parts of continental and oceanic plates. Their different material and temperature, their different viscosity and hydrostatic pressure have been estimated. The difference in viscosity seems to be the most important reason for a number of features which are observed at subduction zones and are consequences of this concept. As mentioned before, large intrusions of viscous lamellas of the oceanic upper mantle intrude less viscous layers of continental crusts and may lead to a number of observed features such as

i. gravity highs behind trenches
ii. peridotitic and gabbroic outcrops at the surface
iii. uplift and forming of horsts
iiii. appearence of double or several Mohos, increase of lamellation.

Possibly, a gap in seismicity should be observed in the lower crust or in zones of low viscosity. As has been mentioned, possibly also a more or less pronounced decoupling between uppermost crust and uppermost mantle takes place in low viscosity zones. This may explain the offset of some mountain roots against their maximum of elevation. It may also play a role for the removing of old roots when mountain ranges erode after the process of compression and subduction has ceased.

The correlation between viscosity and velocity will be extended to more characteristic crustal structures. Especially values of V_S/V_P seem to correlate favourably with the effective viscosity. Results from these studies will be tested by high pressure — high temperature investigations with the specific goal to obtain more exact velocity and viscosity values in the vicinity of melting points. These data should further substantiate our concept of viscosity differences and their influence on geodynamical processes.

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