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46. *Temperature and Electrical Conductivity in the Upper Mantle.*

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Abstract

The temperature in the upper mantle is estimated on the basis of the mineralogical behaviour of mantle materials. Three assumptions are made in the estimation: i.e. (1) Kanamori's Vp-model is the representative of normal upper mantle in a tectonically active region, (2) the upper mantle is composed of olivine-rich peridotite and Fe/Mg ratio of mantle olivine is 1/9, and (3) the cause of the low velocity zone is partial melting. The upper mantle can be divided into two classes in a tectonically active region: (1) hot upper mantle, and (2) normal upper mantle. Two temperature distributions of a different kind are obtained for these mantles. Accordingly, two kinds of electrical conductivity distributions in the upper mantle are determined corresponding to the temperature distributions. The electrical conductivity in the hot upper mantle is two orders of magnitude, at least, higher than that in the normal upper mantle at depths shallower than several tens of kilometers. The high conductivity in the hot upper mantle might possibly correspond to "conductivity anomaly" in the upper mantle. The electrical conductivity thus estimated is lower than 10^{-13} emu even in the anomalous regions, as far as Fe/Mg ratio in mantle olivine [$(Mg, Fe)_2SiO_4$] is assumed to be 1/9. To explain the electrical conductivity higher than 10^{-12} emu as has been estimated in anomalous regions in some cases, the Fe/Mg ratio in mantle olivine should be assumed to be higher than 1/9 (e.g., 1.5/8.5) in the uppermost part of the hot upper mantle, not in the whole upper mantle. The point that a Fe-rich mantle should be assumed in hot regions seems important.

1. Introduction

Existence of large-scale mantle regionality has become evident from seismological investigation (e.g., Katsumata, 1960; Le Pichon et al., 1965; Utsu, 1966, 1967; Oliver and Isacks, 1967; Toksöz et al., 1967)

and from research of terrestrial magnetism (see, Rikitake, 1966a) in recent years. Until recently, seismological evidence has provided most of the knowledge about the interior of the earth. However, recent progress of investigation on transient geomagnetic variations has independently brought out valuable information about the structure of the upper mantle; that is, the recent investigation has established the presence of regional difference in the electrical conductivity (electrical conductivity anomaly) in the upper mantle (Rikitake, 1966a). Such anomalies seem to be closely correlated with seismological regionality.

It is the aim of the present paper to discuss temperature in the upper mantle on the basis of consideration of mineralogical behaviour of the substance composing the upper mantle, and to explain electrical conductivity anomaly in the upper mantle on the basis of the temperature distribution and chemical composition of materials there. In order to do this, it is clear that fairly comprehensive knowledge about the material composing the upper mantle and its mineralogical behaviour is required.

It seems to the writer that the pressure distribution in the mantle has been determined with sufficient accuracy (e.g., Bullen, 1963). As for the substance composing the upper mantle, it seems to have been amply supported that the upper mantle is composed mostly of peridotite or dunite (e.g., Ringwood, 1966a, b).

The temperature distribution within the earth, however, is one of the major unsolved problems in solid earth science. There have been many attempts to estimate the temperature distribution within the mantle and the core, and in these attempts three different methods have been used; i.e. (1) method based on thermal history of the earth (e.g., Jacobs and Allan, 1956; Lubimova, 1958; MacDonald, 1959), (2) method of estimating temperature from physical quantities (such as, seismic wave velocity and electrical conductivity; see, e.g., Verhoogen, 1956; Gilvarry, 1957; Uyeda, 1960), and (3) petrological method (e.g., Verhoogen, 1954; Ringwood, 1966b). However, none of these attempts is free from assumptions on the materials composing the earth and their physical properties. Consequently, the estimated temperature is also affected by the assumptions more or less.

Recent progress in seismology and high-pressure mineralogy has brought out two key points for estimating temperatures in the upper mantle; those are the characteristic velocity gradient at the deeper end of the low velocity zone and the "20° discontinuity." Certain limitations

are to be imposed on temperatures at the respective depths in order to account for their occurrence there. In Section 2, the temperature distribution in the upper mantle will be discussed on the basis of the above two key phenomena.

Since Tozer (1959) and Noritomi (1961) discussed in detail the electrical conductivity in the mantle, the knowledge of electrical conductivity of rock-forming minerals has considerably increased (Bradley et al., 1964; Akimoto and Fujisawa, 1965; Hamilton, 1965; Shankland, 1966; Mizutani and Kanamori, 1967), and some modifications of previous conclusions are needed in the light of the recent experimental results. With the temperature distribution which will be determined in the following section, the writer will in Section 3 estimate a few distributions of electrical conductivity in the upper mantle with the aid of the electrical conductivity data of olivine. The origin of electrical conductivity anomaly will also be discussed in Section 4.

2. Temperature distribution in the upper mantle

The distribution of P-wave velocity provides the most reliable information on the basis of which we may discuss physical properties in the upper mantle. Fine structures of the mantle as displayed by the velocity-depth curves in Fig. 1 have been brought to light by seismological studies in recent years (e.g., Johnson, 1967; Kanamori, 1967). It appears to the writer that there are two interesting features in these models; one is the sharp velocity increase at the bottom of the low velocity zone (will be written as LV zone in the following), and the other the sharp velocity gradient zone at a depth of about 370 km. In this section, discussion will be made on the basis of the velocity-depth curve given by Kanamori (1967), which may represent the curve of a typical island arc tectonic region.

The deeper discontinuity, that is the high velocity gradient zone at the depth of about 370 km, has widely been believed to correspond to the "20° discontinuity" as found in travel-time curves of seismic waves. It is also believed that the zone forms the uppermost part of the transition layer within the earth's mantle, and that such a high gradient is caused by the transition of mantle olivine [mostly $(Mg, Fe)_2SiO_4$] from olivine to spinel structure (e.g., Birch, 1952). Recent experiments on olivine-spinel transition in the Mg_2SiO_4-Fe_2SiO_4 system indicate that the range and the rate of velocity increase observed can possibly be explained

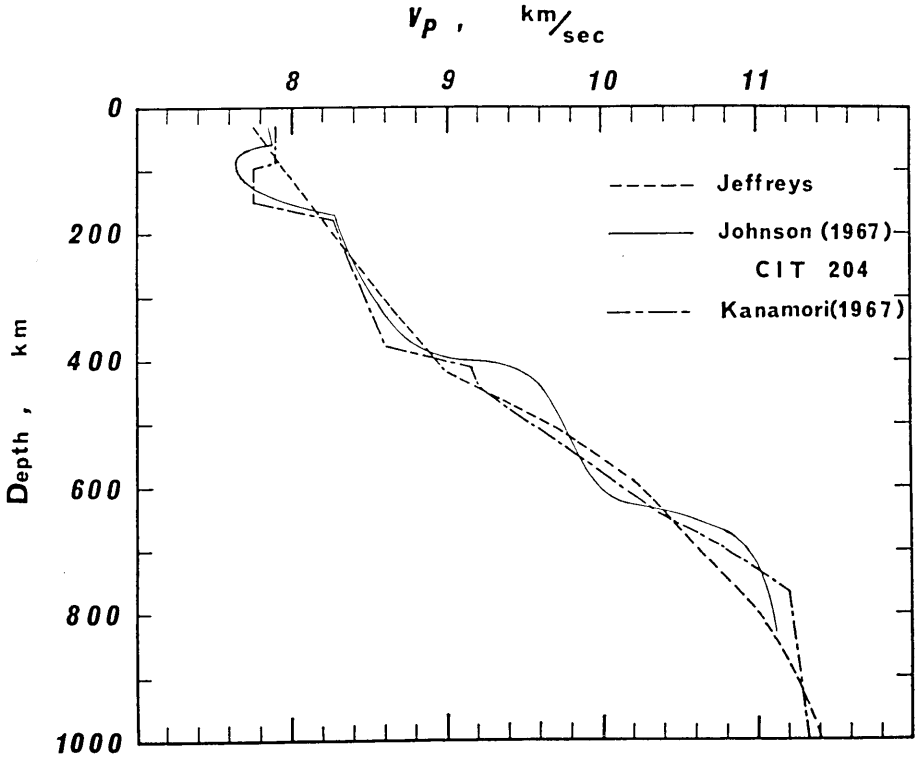

Fig. 1. The latest V_p-models for the mantle after Johnson (1967) and Kanamori (1967) as shown together with Jeffrey's velocity depth curve.

by an olivine-spinel transition of olivines of which the chemical composition may be similar to that of the upper mantle (Akimoto and Fujisawa, 1967; Fujisawa, 1967). In view of such experimental facts, it seems most reasonable to interpret that the discontinuity is given rise to by the olivine-spinel transition. Judging from the phase relation of olivine-spinel transition in the Mg_2SiO_4-Fe_2SiO_4 system, the transition in the actual mantle seems likely to start from a depth of about 350 km and so the discontinuity in the velocity-depth curves by Kanamori and Johnson may be observed at a slightly larger depth (Fujisawa, 1967). If the transition is taken for granted, the experimental results impose a certain limit to the temperature prevailing at that depth. It is thus surmised that the temperature at the depth of 350 km would approximately be in a range between 1000 and 1400°C provided that the Fe/Mg ratio in the mantle-olivine is 1 : 9 (Fujisawa, 1967). It is generally believed that

plausible values for the Fe/Mg ratio of mantle minerals lie between 1 : 9 and 2 : 8, and also that the magnesium-rich side seems more probable (e.g., Harris et al., 1967; White, 1967). Taking experimental errors of the olivine-spinel transition data into account, the writer should think that the most probable value of the temperature at the depth of 350 km in the mantle would be 1400°C (Fujisawa, 1967).

Since the minimum temperature gradient in the earth would be the adiabatic one (Valle, 1951, 1952; Verhoogen, 1951; Shimazu, 1954), the adiabatic temperature distribution which passes the point specified by 1400°C and 350 km on a temperature-depth curve should give the highest possible temperature distribution in the mantle above the depth of 350 km. A usual method of estimating this kind of gradient is to make use of the observed seismic wave velocity combined with the theory of specific heat in *solid*. In the present case the situation is a little complicated, because as described later, the LV zone is interpreted as a partial melting zone in which solid and *liquid* coexist. It is therefore doubtful whether the usual method can successfully be applied to the estimate of the gradient in the LV zone. In the following discussion, the adiabatic gradient is not taken into account for avoiding such an uncertainty. The writer is of the opinion that we had better make use of the line defined by T = 1400°C in place of the adiabatic distribution. The difference between the adiabatic temperature distribution calculated by the classical method (Shimazu, 1954) and the constant temperature line of 1400°C is very small (see, Fig. 5).

In the next place, let us investigate the origin of the characteristic velocity jump which occurs at depths of about 150 to 180 km in the velocity-depth curve given by Kanamori (1967). A number of possible causes of the low velocity zone have so far been presented: (1) high temperature gradients (Birch, 1952; Valle, 1956; Gutenberg, 1959), (2) change in mineralogical composition with depth in the upper mantle (Ringwood, 1962a, b, 1966a, b), and (3) proximity of the melting (Press, 1959) or partial melting (e.g., Anderson, 1967; Kanamori, 1967). As pressure prevailing in this layer can well be attained in laboratories, experimental data may serve for evaluating the above suggestions.

It appears to the writer that the first and second suggestions do not seem plausible as discussed below. We first of all try to explain the velocity jump at about a depth of 160 km, which characterizes the LV zone, assuming only the temperature effect in a homogeneous mantle. The variation of the compressional wave velocity V_p with depth z is given by

$$\frac{dV_p}{dz}=\left(\frac{\partial V_p}{\partial P}\right)_T\frac{dP}{dz}+\left(\frac{\partial V_p}{\partial T}\right)_P\frac{dT}{dz}\,, \tag{1}$$

on the assumption that the upper mantle is chemically homogeneous. If we further assume hydrostatic equilibrium, equation (1) can be written as

$$\frac{dV_p}{dz}=\rho g\left(\frac{\partial V_p}{\partial P}\right)_T+\left(\frac{\partial V_p}{\partial T}\right)_P\frac{dT}{dz}\,, \tag{2}$$

where ρ and g denote the density and the acceleration of gravity respectively. From equation (2), the temperature gradient dT/dz can be expressed as

$$\frac{dT}{dz}=\frac{\dfrac{dV_p}{dz}}{\left(\dfrac{\partial V_p}{\partial T}\right)_p}-\rho g\frac{\left(\dfrac{\partial V_p}{\partial P}\right)_T}{\left(\dfrac{\partial V_p}{\partial T}\right)_P}\,. \tag{3}$$

The upper mantle most possibly being composed of peridotite or dunite, and the pressure and temperature derivatives of the V_p of dunite having already been estimated (e.g., Hughes and Cross, 1951; Birch, 1960, 1961; Simmons, 1964; Kanamori and Mizutani, 1965; Christensen, 1966; see, Table 1), the temperature gradient required to cause the observed velocity jump is determined from relation (3) to be $\cong -15°$K/km, and is shown in Fig. 2 [curve (1), assuming the temperature to be constant at 1400°C at

Table 1. Seismic and material parameters, and $\frac{dT}{dz}$ (for V_p)

model	Kanamori (1967)	CIT 11 CS3*
depth, km	150−180	140−170
$\frac{dV_p}{dz}$, sec^{-1}	1.7×10^{-2}	1.5×10^{-2}
$\left(\frac{\partial V_p}{\partial P}\right)_T$, km/sec kb	$\simeq 2.2\times10^{-2}$ (dunite)** [cf, 1.03×10^{-2} (Mg_2SiO_4)]***	
$\left(\frac{\partial V_p}{\partial T}\right)_P$, km/sec deg.	$\simeq -6.4\times10^{-4}$ (dunite)§ [cf, $\simeq -3.62\times10^{-4}$ (Mg_2SiO_4)]§§	
$\frac{dT}{dz}$, °K/km	$\simeq -15(\simeq -38$†)	$\simeq -13(\simeq -32$†)

* Johnson (1967). ** Mean Value (Birch, 1960, 1961; Simmons, 1964; Kanamori and Mizutani, 1965; Christensen, 1966). *** Schreiber and Anderson (1967). § Hughes and Cross (1951) §§ Soga et al. (1966). † From the data of Mg_2SiO_4.

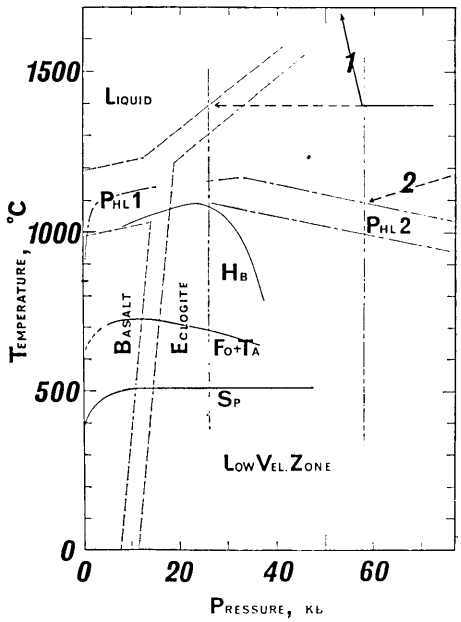

Fig. 2. Mineralogical behavior of basalt and some hydrous minerals: (1) Basalt-eclogite-liquid relation(After Yoder and Tilley, 1962), (2) Dehydration temperatures of some hydrous minerals; serpentine-talc (Sp, Fo+Ta, after Kitahara et al., 1966), hornblende (Hb, after Nishikawa et al., 1968), phlogopite (Phl (1), after Yoder and Eugster, 1954; Phl (2), after Kushiro et al., 1967). Temperature distributions (1) and (2) are also shown.

depths from 180 to 350 km], on the assumption that the upper mantle is composed of dunite alone. In this kind of estimate, the result would not be greatly changed even if more detailed account is taken of the mantle materials. As can be seen in the figure, a temperature drop of several hundred degrees in centigrade within a depth range of a few tens of kilometers is required in order to explain the velocity jump only by a high temperature gradient. Such an enormous temperature drop sounds highly improbable.

Earlier, Valle (1956) estimated the critical temperature gradient for V_p (condition for $dV_p/dz=0$) to be 14°K/km, using simple solid state theory in calculating the effect of pressure and temperature on elastic wave velocities. A similar discussion about the relation between the temperature gradient and shear wave velocity will also be made in the following,

$$\frac{dT}{dz} > -\rho g \frac{\left(\frac{\partial V_s}{\partial P}\right)_T}{\left(\frac{\partial V_s}{\partial T}\right)_P}, \qquad (4)$$

provided hydrostatic equilibrium is present. Birch (1952), and Clark and Ringwood (1964) estimated the critical temperature gradient for V_s from equation (4) to be 5 and 8°K/km, respectively. Valle (1956) also estimated it to be 11°K/km by the similar method for V_p. MacDonald and Ness (1961) examined the critical temperature gradient required for the formation of a region of low velocity as in the following. Since

$$\frac{\left(\frac{\partial V_s}{\partial P}\right)_T}{\left(\frac{\partial V_s}{\partial T}\right)_P} = -\left(\frac{\partial T}{\partial P}\right)_{V_s},$$

equation (4) can be written as

$$\frac{dT}{dz} > \rho g\left(\frac{\partial T}{\partial P}\right)_{V_s}. \tag{5}$$

If the shear wave velocity is a function of density, the righthand-side of (5) can be expressed as

$$\rho g\left(\frac{\partial T}{\partial P}\right)_{V_s} = -\rho g \frac{\left(\frac{\partial T}{\partial \rho}\right)_P}{\left(\frac{\partial P}{\partial \rho}\right)_T} = \frac{\rho g}{\alpha K_T} \approx \frac{g}{\alpha \Phi}, \tag{6}$$

where α is the coefficient of the thermal expansion and K_T the isothermal incompressibility. ϕ is defined as

$$\phi = \frac{K_s}{\rho} = V_p^2 - \frac{4}{3} V_s^2. \tag{7}$$

Taking an approximate expression for the variation of the thermal expansion with pressure, expression (6) can be written as

$$\rho g\left(\frac{\partial T}{\partial \rho}\right)_{V_s} \approx \frac{g}{\alpha_0 \phi}\left(1 + \frac{a(P-P_0)}{K_T}\right), \tag{8}$$

where a is a constant and assumed to be 4 (the value estimated for periclase and for forsterite) in their calculation. The critical temperature gradient estimated by them at various depths shallower than 300 km ranges from 6.51 (at 50 km) to 9.94°K/km (at 150 km) for the Gutenberg's velocity model. Thomsen (1967), who made a similar study, derived the following relation

$$\frac{dT}{dz} = \left(\frac{\partial T}{\partial V_s}\right)_P \left[\frac{dV_s}{dz} - \frac{1+T\alpha\gamma}{\phi} V_s g\left(\gamma - \frac{1}{3}\right)\right], \tag{9}$$

where γ is Grüneisen's ratio. dT/dz in the LV zone is then estimated both for oceanic and continental regions. Used parameters and the results so obtained are listed in Table 2.

Let us examine more simply the temperature gradient for the

formation of the LV zone, in case of V_s by using the relation similar to (3)

$$\frac{dT}{dz}=-\rho g\frac{\left(\frac{\partial V_s}{\partial P}\right)_T}{\left(\frac{\partial V_s}{\partial T}\right)_P}+\frac{\frac{dV_s}{dz}}{\left(\frac{\partial V_s}{\partial T}\right)_P}, \tag{10}$$

provided a chemically homogeneous mantle is assumed. Data for various quantities used in the estimate are listed in Table 2 along with the results. If we adopt a value of $T=300°C$ at 35 km, and gradients of 20°C/km for depths of 35 to 90 km (oceanic, see Table 2) and 10°C/km from 90 to 150 km (the critical gradient), the temperature at a depth of 150 km will be 2000°C beneath the oceanic region, provided the high temperature gradient were assumed to be solely responsible for the LV zone. Such a high temperature would inevitably cause complete melting

Table 2. Seismic* and material parameters, and $\frac{dT}{dz}$ (for Vs)

model	oceanic	continental
	CIT 11†	Gutenberg††
LV zone, km	35−90*	45−140*
$\frac{dVs}{dz}$, sec^{-1}	-4.9×10^{-3}*	-2.2×10^{-3}*
ρ, g/cm^3	3.35*	
g, cm/sec^2	980*	
$\frac{dT}{dz}$, °K/km	9.5−28*	4.8−19*
$\left(\frac{\partial Vs}{\partial P}\right)_T$, km/sec kb	$\simeq1.3\times10^{-2}$ (dunite)** [cf, 2.4×10^{-3} (Mg_2SiO_4)]***	
$\left(\frac{\partial Vs}{\partial T}\right)_P$, km/sec deg	$\simeq-4.8\times10^{-4}$ (dunite)§ [cf, $\simeq-2.71\times10^{-4}$ (Mg_2SiO_4)]§§	
$\frac{dT}{dz}$, °K/km	≃20(≃21†††)	≃14(≃11†††)

* Seismic parameters and (dT/dz) in upper lines are reproduced from Table 1 and 4 in Thomsen (1967). † Anderson and Toksöz (1963). †† Gutenberg (1959). ** Mean value (Birch, 1960, 1961; Simmons, 1964; Kanamori and Mizutani, 1965. Christensen, 1966). *** Schreiber and Anderson (1967). § Birch (1943). §§ Soga et al. (1966). ††† From the data of Mg_2SiO_4

of the upper mantle. This completely contradicts observational evidence. Accordingly it is necessary to look for another cause of the LV zone; that is, vertical inhomogeneity—changes in the mineralogical composition with depth or partial melting—in the upper mantle.

Let us in the next place examine the possibility of accounting for the LV zone by taking into account changes in the mineralogical composition with depth. The changes must be caused by a vertical zoning in chemical composition and also by the changes in the pressure and temperature. In this case the widespread presence of a particular substance which is characterized by a low elastic wave velocity should be admitted in the LV zone. There are two suggestions for this problem; (1) suggestion represented by the "pyrolite model" (Ringwood, 1962a, b; Green and Ringwood, 1963, 1967a), (2) suggestion of the presence of hydrous minerals, which necessarily lowers wave velocity in the LV zone (e.g., Oxburgh, 1964). As is discussed below, however, all these suggestions fail in the end to explain the presence of the LV zone, and partial melting hypothesis will finally be introduced.

For the pyrolite model, occurrence of the LV zone may be attributed to mineral reactions in minor constituents of "pyrolite". Material of pyrolite composition, which is equivalent to a mixture of three parts of dunite or peridotite to one part of basalt, could crystallize in four distinct mineral assemblages in the range of temperatures and pressures occurring in the upper mantle (Green and Ringwood, 1963; modified, 1967a). In the case of the modified model, the LV zone is divided into three mineralogically different layers; that is, from shallower depths, (1) olivine ($\sim$50%)-pyroxene($\sim$45%)-spinel ($\sim$5%) layer [30–70 km, $dT/dz\sim$ 13°K/km-from Ringwood's (1966b) oceanic geotherm], (2) olivine ($\sim$55%)-aluminus pyroxene ($\sim$40%)-garnet ($\sim$5%) layer [70–150 km, $dT/dz\sim$ 5.5° K/km], (3) olivine ($\sim$55%)-pyroxene (40–35%)-garnet (5–10%) layer [150–190 km, $dT/dz\sim$2°K/km], and below this zone there exist olivine ($\sim$55%), pyroxene ($\sim$35%), and garnet ($\sim$15%). As is seen from the above, the variation in the mineral assemblages throughout the upper mantle is not so large, and the difference in seismic wave velocity between these assemblages may be negligible because the elastic wave velocities of olivine are approximately the same as those of garnet, while the velocities of pyroxene differ very little from them (see e.g., Birch, 1960, 1961; Christensen, 1965). The upper mantle, therefore, can be treated approximately as homogeneous in this model. From the discussion earlier in this section, it is clear that the LV zone shown in the recent

velocity models can not occur in a homogeneous mantle by the temperature effect only. Since the pyrolite model seems to fail, we will next consider the possibility of the presence of hydrous minerals in the *LV* zone.

There is petrological evidence that serpentine, talc, amphiboles and micas are likely to be present in the upper mantle (e.g., Kushiro, 1966).

Table 3. Seismic and material parameters, and volume content of hydrous minerals and liquid

I. V_p (at 10 kb, room temperature)

minerals	V_p(km/sec)	reference	representative value (km/sec)
Serpentine			
	6	7	
	5.2	6	6
	6.97	8	
serpentinite	6.00-6.82	3	
Talc			
talc schist	7	3	7
talc schist	6.97	4	
Amphibole			
amphibolite	7.35	3	
amphibolite 1	6.9*	5	
amphibolite 2	6.8*	5	7
hornblende I	6.81*	1	
hornblende II	7.04*	1	
actinolite schist	7.54	3	
Mica			
muscovite	5.78*	2	
biotite	5.26*	2	5.5
phlogopite	5.55*	2	
Dunite			
	7.78	3, 4, 8	
	8.00	3, 4, 8	
	8.28	3, 4	8.3
	8.52	7	
	8.42	3, 4, 8	
	8.66	6	

(to be continued)

(Table 3, continued)

II. Seismic parameters, and volume content of hydrous minerals and liquid

model	Vp		Vs	
	Kanamori**	CIT 11 CS3***	CIT 11§	Gutenberg§§
depth, km	150-180	140-170	35-90	45-140
ΔV, km/sec	0.5	0.5	0.25	0.25
Volume Content, %				
serpentine+talc	20-30			
amphibole	35			
mica	17			
Vp_0, km/sec	8	8		
Vs_0, km/sec			4.5	4.5
volume content of liquid, %	14	14	5	4

References
1. Alexandrov and Ryzhova (1961a)
2. Alexandrov and Ryzhova (1961b)
3. Birch (1960)
4. Birch (1961)
5. Christensen (1965)
6. Christensen (1966)
7. Kanamori and Mizutani (1965)
8. Simmons (1964)

* Mean value of Voigt and Reuss average (estimated for 1 atm, see, Christensen, 1965)
** Kanamori (1967)
*** Johnson (1967)
§ Anderson and Toksöz (1963)
§§ Gutenberg (1959)

The compressional wave velocities of these hydrous minerals are listed in Table 3. The elastic wave velocity in bulk of mixture of dunite and hydrous minerals is calculated both by Voigt average and by Reuss average (see, e.g., Huntington, 1958), and mean values are used for the estimate of amount of each hydrous mineral required for the formation of the LV zone. The estimation has been made on the assumption that the LV zone is composed both of dunite and one of the hydrous minerals mentioned above (serpentine and talc always make a pair; see, discussion below), and also that there exists dunite only above and below the zone. Results also are listed in Table 3.

All these hydrous minerals can only exist under a limited pressure-temperature condition. They are all dehydrated at higher pressures and temperatures (see, Fig. 2). Now we examine the possibility of the existence of them in the *LV* zone. Phase relation in the system MgO-SiO_2-H_2O, which is closely related to dehydration of serpentine and talc, is extensively investigated (Kitahara et al., 1966; Kitahara and Kennedy, 1967). Among many complicated reactions in this system, the rections

$$
\begin{array}{lll}
 & \text{(i)}\ \ \underline{\text{FORSTERITE}+\text{ENSTATITE}+H_2O} & \\
 & \qquad\qquad\qquad\qquad \downarrow\uparrow \text{(a)} & \\
 & \text{(ii)}\ \ \underline{\text{FORSTERITE}+\text{TALC}+H_2O} & \\
 & \downarrow\uparrow \text{(b)} \qquad\qquad \downarrow\uparrow \text{(c)} & \\
\text{(iii)}\ \ \underline{\text{SERPENTINE}+\text{TALC}} & & \text{(iv)}\ \ \underline{\text{SERPENTINE}+\text{FORSTERITE}+H_2O} \\
 & & \qquad\qquad\qquad \downarrow\uparrow \text{(d)} \\
 & & \text{(v)}\ \ \underline{\text{SERPENTINE}+\text{BRUCITE}}
\end{array}
$$

are of great importance in the ultrabasic upper mantle. Since there are no hydrous minerals in the mineral assemblage (i), reaction (a) is a final dehydration in these mineral assemblages. The *P-T* conditions for the reactions (a), (b) and (c), determined by Kitahara and others (1966), are shown in Fig. 2. The reaction temperatures of (b) and (c) are equal. As is seen in the figure, serpentine and talc are unable to exist at temperatures above 700°C under pressures prevailing in the upper mantle. Since the V_p-model of Johnson or Kanamori, to which the present discussion is referred, may represent the character of a tectonically active region, where basaltic magma could be produced, our discussion is concerned mostly with this region. As will be discussed later, temperatures much higher than 700°C being required for producing basaltic magma, the temperature should be higher than 700°C there, and no serpentine and talc may be present at depths below several tens of kilometers in this region.

Oxburgh (1964) suggested the presence of amphibole in the upper mantle as a potassium-bearing mineral and also as a possible explanation for the *LV* zone. However, thermodynamical calculations show that amphiboles would probably break down to anhydrous high pressure mineral assemblage plus H_2O in relatively shallow depths (perhaps, shallower than 100 km) of the upper mantle (Kushiro, 1966; Green and Ringwood, 1967b). Recent high-pressure experiments on a natural hornblende (common hornblende, Fe/Mg=10/16) also show that this hornblende

can be stable only in the shallower depths (less than 100 km) in the upper mantle even if water pressure is equal to total pressure (Nishikawa, 1968; see, also, Nishikawa et al., 1968; Fig. 2). Amphibole, therefore, would not be able to exist in the *LV* zone.

As for micas, data of dehydration are insufficient for detailed discussion. Under relatively low pressures (less than 3 kb) most micas are dehydrated at temperatures below 900°C (see, e.g., Clark, 1966). Hence most of them would possibly break down to anhydrous minerals plus H_2O at relatively low temperatures. Among common mica minerals, phlogopite would be most stable under high pressures prevailing in the upper mantle (Williams, 1932; Kushiro, 1968). Accordingly, high-pressure experiments have made on the stability of natural and synthetic phlogopites by several authors. Yoder and Eugster (1954) examined it in a lower pressure range (less than 5 kb). Experiments on natural phlogopite by Markov and others (1966) show that phlogopite breaks down to garnet and unidentified minerals at 66 kb at temperatures below about 1500°C. Kushiro and others (1967) investigated the stability condition for natural and synthetic phlogopites in an extensive range of pressure and temperature. These results are reproduced in Fig. 2. If the presence of phlogopite is responsible for the formation of the *LV* zone, the stability condition of phlogopite must be satisfied there; that is, the temperature at the depth of 180 km should be 1100°C at most, and accordingly the temperatures at depths shallower than 180 km must be lower than 1100°C. Under these temperatures, the stability conditions of both phlogopite (and perhaps some other kinds of micas) and amphiboles are automatically satisfied from the top to a depth of 180 km in the upper mantle (see, Fig. 2). As far as we assume that the *LV* zone is formed by the presence of hydrous minerals, they must be concentrated only in the *LV* zone. This requires another mechanism of their concentration; for instance, concentration of several kinds of elements (Na, K, Ca, and Al), or H_2O, or both. Although the possibility of such a concentration of elements and water cannot be ruled out, the writer thinks it highly improbable. Widespread concentration of hydrous minerals in the *LV* zone, therefore, seems unlikely.

The only remaining possibility for the cause of the *LV* zone is the occurrence of partial melting. Occurrence of partial melting in the *LV* zone has been suggested from recent seismological evidence (Anderson, 1967; Kanamori, 1967). The elastic wave velocities of solid mixed with a small fraction of liquid are approximately expressed as

$$\left.\begin{aligned} V_p &= V_{p0}\left(1 - \frac{4}{9}c\right) \\ V_s &= V_{s0}(1 - c) \end{aligned}\right\} \tag{11}$$

where V_{p0} and V_{s0} are the compressional and shear wave velocities of host solid, and c is the volume content of the melted portion (Hashin, 1962; Kanamori, private communication). The volume content of melted portion necessary for giving rise to a sharp velocity change around 160 km in the present two V_p-models and also for causing the decrease in V_s in the two velocity models is calculated from equation (11), using parameters listed in Table 3. Estimated values are also listed in the same table. We see in the table that liquid contents from 4 to 14 per cent can result in conceivable velocity values.

Melting point of basalt (e.g., Yoder and Tilley, 1962; Cohen et al., 1967) and peridotite (e.g., Ito and Kennedy, 1967; Green and Ringwood, 1967a) under pressures have been extensively investigated (see, Fig. 2). It has been suggested in this paper that the temperature in the mantle shallower than 350 km should not exceed 1400°C. The peridotite solidus under pressures higher than about 26 kb (below about 85 km depth) lies substantially greater than 1400°C (see, Fig. 2). Occurrence of partial melting of peridotite, therefore, seems impossible in the LV zone.

Meanwhile, it has been reported that existence of a minute amount of water greatly lowers the melting point of minerals (e.g., Yoder and Tilley, 1962; Shimada, 1966; Kushiro, 1966; see, Fig. 3). As is seen in Fig. 3, the peridotite solidus lies below 1400°C at depths shallower than 180 km in case some water exists in the upper mantle. Occurrence of partial melting may be possible in the LV zone in this case. It has been generally accepted by geologists and geophysicists that the primary earth is homogeneous with primordial chemical composition, and that surface water has possibly been squeezed out from the interior of the earth all through geologic times. It would not be unreasonable to suppose that some water might be trapped in the upper part of the mantle during the physico-chemical development; that is, in hydrous minerals, as crystal water, and around grain boundary as free water. It is widely believed that basaltic magma produced in the mantle contains water, because the chemical composition of its solidification products (such as, basalt, gabbro, and dolerite) usually contains H_2O. Peridotite inclusions in basaltic magma, which are believed to be fragments of mantle materials, sometimes contain amphiboles (Lausen,

1927; White, 1966; Kuno, 1967), and rarely micas (Wilshire and Binns, 1961). In kimberlite pipes phlogoptite occurs in both kimberlites and eclogite inclusions (Williams, 1932). Some parts of H_2O in these rocks, of course, are due to contamination or alteration. Some portion of it, however, may be of mantle origin. Although no evidence for quantitatively estimating the content of H_2O in the upper mantle has been presented, the existence of water there seems highly likely. In the following discussion the existence of H_2O in the upper mantle is assumed.

The solidus of peridotite (lherzolite)-water system (Kushiro et al., 1968) are shown in Fig. 3. In the present model, the temperatures in the *LV* zone must be higher than the peridotite-water solidus (or the

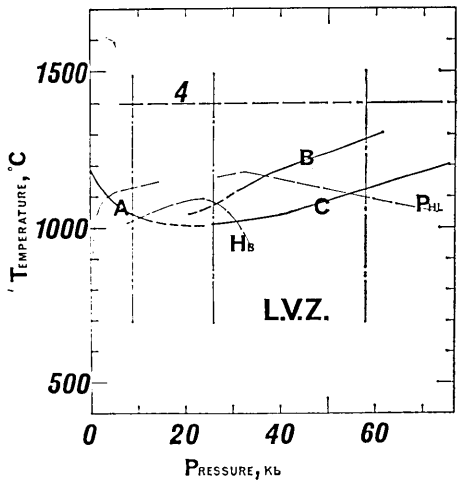

Fig. 3. Melting temperature of basalt and peridotite under high water pressures: Liquidus of basalt-water system, (A)-Shimada (1966), (B)-Nishikawa (1968); Solidus of peridotite-water system, (C)-Kushiro et al. (1968). Temperature distribution (4), and dehydration temperatures of hornblende (H_B) and phlogopite (P_{HL}) are also shown.

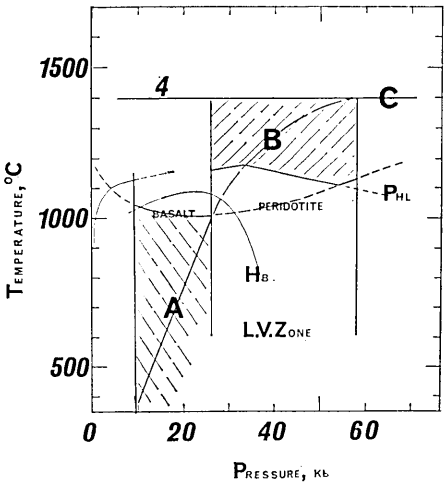

Fig. 4. Permitted range of temperature in and above the low velocity zone (hatched area), and the temperature distribution being estimated in accordance with this area. Temperature distribution (4), and dehydration temperatures of hornblende and phlogopite are also shown.

basalt-water liquidus, if it coincides with the peridotite-water solidus), and the temperatures above the *LV* zone must be lower than the peridotite-water solidus (or basalt-water liquidus). The liquidus curves of basalt-water system in three kinds of typical basalts (tholeiite, high-alumina, and alkali) are in good agreement at pressures lower than 10 kb (Yoder and Tilley, 1962). No experiment on peridotite-water system in

this pressure range has been made yet. Complication arises in the LV zone. As is seen in Fig. 3, liquidus of basalt (tholeiite, natural)-water system (Nishikawa, 1968) is greatly different from the peridotite-water solidus. Such a difference may be of great importance for discussing magma generation by partial melting of peridotite. However, the most important point of the present discussion about the possibility of forming the LV zone is the existence of liquid, of which the chemical composition may or may not be basaltic. Accordingly, the peridotite-water solidus only is taken into account. This solidus is substantially lower than 1400°C. If there were sufficient water in and below the LV zone, partial melting might extensively occur there. This is inconsistent with observational evidence. Since basaltic magma produced in the upper mantle is not saturated with water, the H_2O content seems very small in the upper mantle except in its uppermost part (Ringwood, 1966b). Very little free water may, therefore, be included in the LV zone and probably non-existing below this zone. If the amount of H_2O in the LV zone were very small and the temperatures were lower than the dehydration temperature, the water might possibly be consumed to form hydrous minerals. In order to have free water and partial melting in the LV zone, the temperatures there must be higher than the dehydration temperatures of hydrous minerals existing in this zone.

Thus the permitted range of temperatures (hatched area in Fig. 4) in and above the LV zone can be determined as shown in Fig. 4. Let the temperature be constant at 1400°C between 180 and 350 km [(C) in Fig. 4]. The temperature immediately above the LV zone must be lower than 1000°C (see, Fig. 4). As is seen in Fig. 1, V_p is nearly constant at depths between *moho* and the LV zone in the velocity models to-date. The critical temperature gradient, under which V_p is constant, is determined to be $\simeq 12°$ K/km, making use of the pressure and temperature derivatives of elastic wave velocity of dunite (see, Table 2). Temperature distribution (A) in Fig. 4 is determined on the basis of the critical gradient and passes the point specified by 1000°C and 85 km. The distribution would be the highest possible one above the LV zone, as far as Kanamori's Vp model is considered. No other restriction on temperatures exists than as shown in Fig. 4 in the LV zone, and the temperature distribution (B) is drawn arbitrarily to connect (A) with (C).

All the temperature distributions referred to in this paper are shown in Fig. 5. Curve (3) in this figure corresponds to the highest possible temperature distribution in a tectonically active region, where the V_p-depth curve is similar to Kanamori's model. The velocity distribution above the

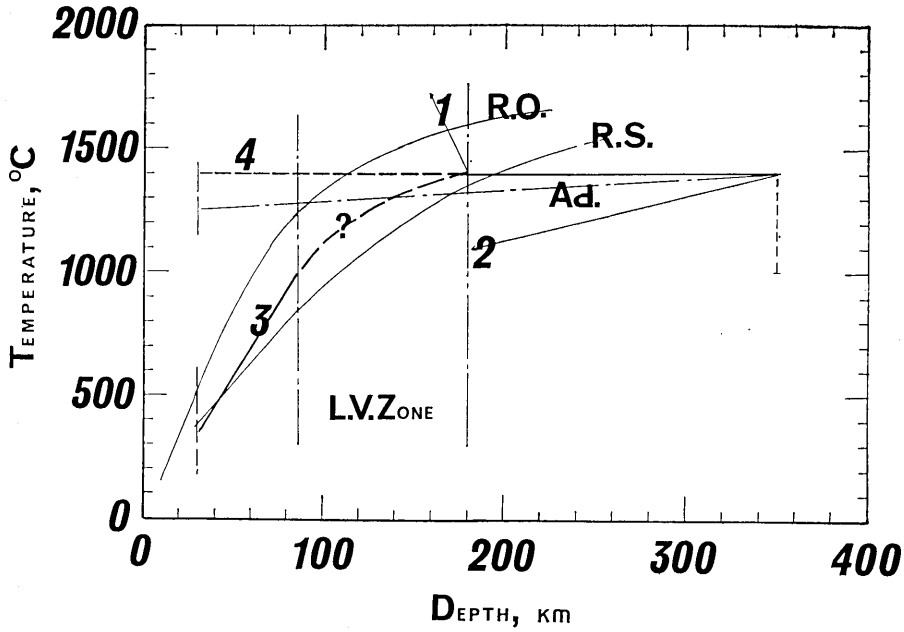

Fig. 5. All temperature distributions discussed in Section 2 as shown together with the adiabatic temperature distribution (Ad. Shimazu, 1954) and Ringwood's geotherm (Ringwood, 1966b; R.O.-oceanic, R.S.-shield).

LV zone may be greatly different from Kanamori's model in the region where temperatures are similar to branch (4). The *LV* zone would extend upward as far as to *moho* in such a region, and anomalous features in attenuation of seismic waves, electrical conductivity, or other physical properties could possibly be observed. Consequently the upper mantle can be divided into two classes: (A) hot upper mantle, where the temperature distribution is similar to (4) in Fig. 5, and (B) normal upper mantle, where the temperature distribution is similar to (3). As is seen in this figure, the adiabatic temperature distribution, of which the gradient is estimated by a classical method (Shimazu, 1954), deviates very little from 1400°C all through the upper mantle.

Partial melting in the *LV* zone as discussed in this section may also have a bearing on magma generation.

3. The electrical conductivity in the upper mantle

Since Tozer (1959) and Noritomi (1961) investigated the electrical conductivity in the mantle, considerable progress in experiments on electrical conductivity of rock-forming minerals, especially of olivine

(Bradley et al., 1964; Akimoto and Fujisawa, 1965; Hamilton, 1965; Shankland, 1966; Mizutani and Kanamori, 1967), has been made. It appears to the writer that some modifications of previous conclusions are needed in the light of the recent experimental results. It is the aim of this section to estimate the electrical conductivity in the upper mantle on the basis of the temperature distributions determined in the former section and the most recent data of the electrical conductivity of olivine.

It is assumed that the main constituent of the upper mantle is olivine-rich peridotite, and also that Fe/Mg ratio of olivine [$(Mg, Fe)_2 SiO_4$] in the upper mantle is 1/9 throughout this paper, so that the electrical conductivity of upper mantle materials is assumed to be represented by that of olivine in the following discussion. The electrical conductivity of olivine can be approximately represented by the expression

$$\sigma = \sigma_0(P, m, n_i) \exp\left(-\frac{E(P, m, n_i)}{kT}\right) \tag{12}$$

for the temperature range prevailing in the upper mantle, where P and m specify pressure and bulk chemical composition (Fe/Mg ratio) respectively, n_i specifies impurities or micro structures, and σ_0 and E depend on all of them. E is activation energy, k Boltzmann's constant, and T absolute temperature.

The effect of pressure on the conductivity is so small that it is hardly determined with the present accuracy of experiment (Akimoto and Fujisawa, 1965; Hamilton, 1965), so that no account of the pressure effect will be taken in the following. As the Fe/Mg ratio is assumed as 1/9, experimental data of which the ratio is close to this value will be chosen for estimating the mantle conductivity. The effect of impurities or micro structures on the conductivity of olivine has not yet been well established, but the presence of Fe^{3+} ion in olivine would be the most important as shown by Bradley and others (1964). They showed a systematic relation between content of ferric iron and electrical conductivity in fayalite. A small amount of Fe^{3+} ion may replace Fe^{2+} ions in olivine lattice, and its quantity may be determined by the oxygen partial pressure at its genesis. Magnetite coexisting as impurity, which is easily detected in many cases if examined closely, sometimes causes a considerable error in the electrical conductivity measurement. Olivines synthesized or measured under an unsuitable atmospheric condition may be oxidized and yield magnetite. There exist rarely small amounts of magnetite, chromium-rich magnetite, and chromite in natural olivine

(Deer et al., 1962). Samples for measurement of the electrical conductivity, therefore, must be carefully examined, and stability conditions must be satisfied through the measurement. Electrical conductivity of olivines generally increases with increase in impurity content. Reliability of data may be checked by examining their reproducibility and by comparison of the conductivity values of olivines having the same Fe/Mg ratio.

Most of the recent results of the measurements on olivines are compiled in Fig. 6. Samples synthesized by Shankland (1966) seem most reliable, but his measurements are limited only to single crystals of olivines of compositions similar to forsterite (0.05 and 0.2% fayalite) at ordinary pressure. The data by Bradley and others (1964) are rather confusing. The relation between the electrical conductivity of $(Mg_{0.5}Fe_{0.5})$-olivine and those of the three fayalite samples with Fe^{3+} *ion* content different from one another is complicated. The conductivity of $(Mg_{0.9}Fe_{0.1})$-olivine reported by them is considerably higher than that of natural olivine, having nearly the same Fe/Mg ratio, as measured under the similar pressure (Hamilton, 1965). Even in olivines with the same Fe/Mg ratio, their electrical conductivities usually do not agree with each other. Here, the writer tentatively chooses the lowest of them as a typical conductivity versus temperature relation in the upper mantle, because existence of impurity and other secondary elements generally causes an increase of the electrical conductivity of olivines. At present, therefore, we use Hamilton's (1965) data of olivine, containing 10.4 per cent fayalite, measured at 21.6 kb. It should be borne in mind, however, that the conductivity-temperature relation thus adopted is only tentative. Although occurrence of partial melting is suggested in the *LV* zone, it has been shown by Tozer (1959), Rikitake (1962) and others that the molten portion, 10 per cent or so in volume content, affects very little the conductivity in bulk.

We are now in a position to estimate the electrical conductivity distributions in the upper mantle on the basis of the temperature distributions determined in the previous section. The assumptions necessary for such work are summarized below: (1) The upper mantle is composed of olivine, Fe/Mg ratio of which is 1/9, (2) pressure effects on the electrical conductivity are negligible, (3) Hamilton's data of natural olivine containing 10.4 per cent of fayalite, measured at 21.6 kb, is the representative of the mantle-olivine, (4) and the effect of partial melting on the conductivity can be ignored. Results so obtained are shown in Fig. 7. Numbers attached on each curve in the figure correspond to those attached to each temperature-depth curve in Fig. 5, from which the

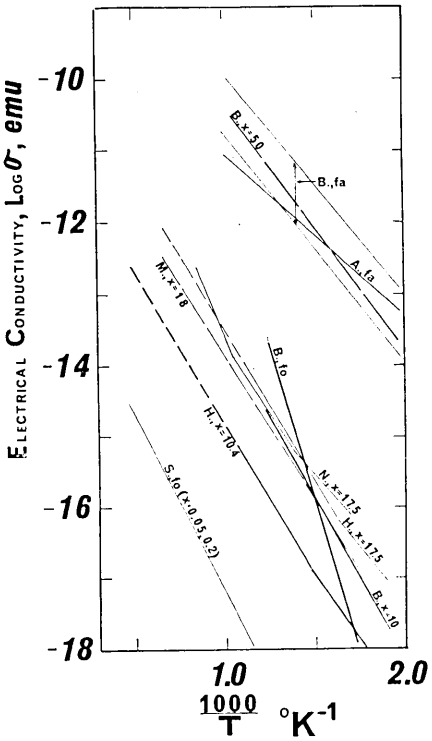

Fig. 6. Electrical conductivities of olivine [$(Mg, Fe)_2SiO_4$] as recently determined.

A.-Akimoto and Fujisawa (1965), 31 kb
B.-Bradley et al. (1964), 23.6 kb.
H.-Hamilton (1965), 21.6 kb.
M.-Mizutani and Kanamori (1967), 1 atm.
N.-Noritomi (1961), 1 atm.
S.-Shankland (1966), 1 atm.
fo-forsterite (Mg_2SiO_4).
fa-fayalite (Fe_2SiO_4).

$$x=\frac{fa}{fo+fa}\times 100$$

conductivity distribution is estimated.

4. Explanation of conductivity anomalies

Recent investigations on geomagnetic variation strongly suggest the presence of regional high conductive area—conductivity anomaly—in the upper mantle (see, Rikitake, 1966a). Probably typical profiles of conductivity anomaly may be those which are schematically shown in Fig. 8.

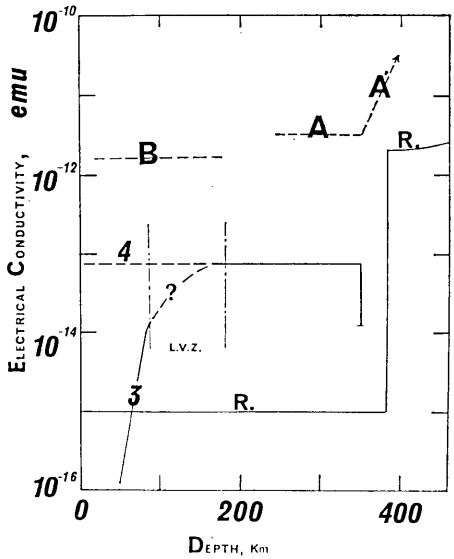

Fig. 7. Electrical conductivity distributions in the upper mantle as estimated in the present paper. Rikitake's distribution (Rikitake, 1950a, b, c; R.) is also shown.

The anomalies can possibly be divided into two types from configuration of high conducting materials; that is, crustal (surface)-layered anomaly (e.g., Sierra Anomaly, Schmucker, 1964; see, II in Fig. 8), and wedge or step type anomaly (e.g., Alert Anomaly, Rikitake and Whitham, 1964; Texas Anomaly, Schmucker, 1964; California Coast Anomaly,

Schmucker, 1964; Filloux, 1967; see, I, III, and IV in Fig. 8).

Crustal anomaly, which will not be discussed here, may be due mainly to the presence of a high conductive sedimental layer. In the case of a wedge or step type anomaly, a considerably large contrast in the electrical conductivity is present between the anomalous region and the neighbouring normal region. The causes of the contrast may be attributed to the difference in temperature between them, or to the lateral chemical inhomogeneity, or to both. Two different distributions [(3) and (4) in Fig. 7] of electrical conductivity may be obtained corresponding to the temperature distributions (3) and (4) in Fig. 5. As is seen from Fig. 7, the difference amounting to two orders of magnitude between two regions of different temperature distribution may possibly be present at depths shallower than several tens of kilometers. Regions, where the electrical conductivity distribution is similar to (4) in Fig. 7, might correspond to the anomalousareas—"conductivity anomaly." The electrical conductivity in these regions is estimated to be about 10^{-13} emu, using the Fe/Mg ratio of 1/9 (see, Fig. 7).

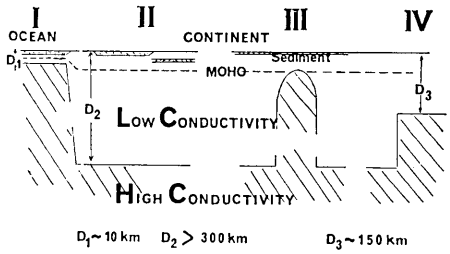

Fig. 8. Schematic representation of typical profiles of conductivity anomaly in the crust and the upper mantle.

It often provides a good approximation to assume the electrical conductivity in an anomalous region simply as infinity. Such a perfectly conducting model usually suggests conductivity values higher than 10^{-12} emu. It seems likely, therefore, that the electrical conductivity in the anomalous regions is considerably higher than 10^{-13} emu [(4) in Fig. 7], though no accurate determination of the conductivity is present at the present stage of the investigation. If such a high conductivity ($\gg 10^{-13}$ emu) is present in relatively shallow depths in the upper mantle, it may be attributed to higher temperature [than (4) in Fig. 5], or higher Fe/Mg ratio ($>1/9$), or other unknown mechanisms, or their combined effect. The most plausible interpretation is the presence of anomalously higher temperatures, or higher Fe/Mg ratio, or both in the anomalous regions.

Let us now examine the possibility of the presence of higher temperature without change of the assumption of the Fe/Mg ratio as 1/9. The maximum possible temperature in the upper mantle is closely related to the depth of occurrence of seismic discontinuity around 350 km as

previously discussed. In most of the recent V_p-models the seismic discontinuity occurs at depths shallower than that in the model of Kanamori or Johnson (see, Johnson, 1967). The occurrence of the discontinuity at shallower depths suggests temperatures rather lower in the upper mantle than that estimated in the previous section, as far as Fe/Mg ratio of mantle olivine is kept at 1/9. It is therefore necessary to look for other causes of the regional high conductivity in the upper mantle.

It is widely believed that the Fe/Mg ratio in mantle olivine would not exceed 2/8. Let us consider two different models with higher Fe/Mg ratio; (A) Fe/Mg ratio of olivines is 1.5/8.5 throughout the upper mantle, and (B) only at relatively shallow depths in the anomalous regions, where the temperature distribution is similar to (4) in Fig. 5, Fe/Mg ratio of olivine is 1.5/8.5, and the ratio in the remaining region is 1/9.

In model (A) the temperature at the depth of 350 km should be raised to 1600°C in order to have the olivine-spinel transition there (Fujisawa, 1967). Consequently, the electrical conductivity is higher than 10^{-12} emu [see, (A) in Fig. 7; estimated using the data of (17.5% Fa)-olivine, Hamilton, 1965] even at a depth immediately above the level of the transition, and the electrical conductivity may be considerably higher than 10^{-11} emu just below this depth because of the occurrence of the olivine-spinel transition there (Akimoto and Fujisawa, 1965) [see, (A) and (A′) in Fig. 7]. This completely contradicts Rikitake's estimation of the electrical conductivity in the mantle on the basis of geomagnetic analyses (Rikitake, 1950a, b, c; see, also Rikitake, 1966b; see, R. in Fig. 7). It is difficult, therefore, to explain conductivity anomaly by this model.

In model (B), the temperature indicated by distribution (4) in Fig. 5 may not be changed in the anomalous regions. The electrical conductivity in this case is shown in Fig. 7 [curve (B)]. The electrical conductivity for the corresponding normal region is also curve (3) in Fig. 7. An electrical conductivity higher than 10^{-12} emu seems possible for the model at depths of a few tens of kilometers in the anomalous region (see, Fig. 7). Seismic wave velocity in this region should be very low, and high attenuation of seismic waves would also be observed there. The model (B)—slightly higher concentration of iron in hot upper mantle—is highly probable for the cause of a regional high conductivity in the upper mantle—"conductivity anomaly."

Exact values of the Fe/Mg ratio in mantle materials, however, can not be determined at the present stage of the investigation, though it seems likely that the ratio amounts to nearly 1/9 (e.g., Harris et al.,

1967; White, 1967). Discussion about its regional variation is even more difficult.

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46. 上部マントルの温度と電気伝導度

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最近の，高温高圧実験から得られた，岩石や鉱物の，物理的性質や，鉱物学的性質に関するデータと，地震学的研究によつて得られた，上部マントルに関する諸事実とを，組合せて，上部マントルの温度を推定した．さらに，推定された温度分布と，最近得られた造岩鉱物，特に，オリビンの，電気伝導度のデータとを用いて，上部マントルの電気伝導度の分布を推定し，さらに，上部マントルの，電気伝導度異常について，その成因を考えた．

本研究において，用いられた仮定，および，得られた結論は，下記の通りである．

〔仮定〕

1. マントルは，主として，かんらん岩で，構成されていると考える．また，かんらん岩の主成分であるオリビンの，Fe/Mg 比は，1/9 と仮定する．

2. 金森により得られた，マントルの Vp 分布は，活動的な造構帯の，標準のマントルを代表するものであると仮定する．この場合，マントルの，深さ約 350 km での温度は，1400°C 以下である．

3. 上部マントルの，地震波の低速度層は，マントルの物質の部分溶融によると考える．

〔結論〕

1. 造構帯およびその周辺においては，上部マントルは，2つの種類に分類できる．即ち，(1) 高温帯 (hot upper mantle) と，標準帯 (normal upper mantle) である．つまり，この2領域に対応して，2種の温度分布が存在する．そしてこの2種の温度分布は，互に著しく異なつている．特に，高温帯では，地震波速度の分布は，金森のモデルとは，著しく異なつているはずであり，また，地震波の減衰も著しいはずである．さらに，この2種類の温度分布に対応して，2種類の上部マント

ルにおける電気伝導度の分布が得られる．高温帯での，電気伝導度は，標準帯に比し，マントルの数十km より浅い所では，2桁以上大きい．これが，いわゆる上部マントルの〔電気伝導度異常〕に相当する可能性は極めて高い．

2. マントル-オリビンの Fe/Mg 比を 1/9 とする限り，上部マントルの電気伝導度の上限は，たとえ高温帯においても，10^{-13}emu 程度である．

3. 一方，いくつかの伝導度異常帯では，その電気伝導度が，10^{-12} emu 以上であると，報告されている．このような高伝導度を説明できる最も可能性のあるモデルは，異常帯(高温マントルの最上部)で，マントル-オリビンの Fe/Mg 比が，1/9 より少し大きくなつている (例えば，1.5/8.5) と考えることである．この場合，異常帯では，電気伝導度は，10^{-12} emu より，かなり大きくなる．上部マントル全域にわたつて，Fe/Mg 比を増すと 350～400 km の，地震波の不連続を，オリビン-スピネル転移によると考える限り，深さ 350 km での温度は，先に述べた，1400°C より著しく高くなるはずである．したがつて，電気伝導度の増大する割合は，さらに大きくなり，地磁気変動の解析から推定されている，マントルの電気伝導度の分布の様相とは著しく矛盾することになる．