of conductivity of natural olivine samples with cycling the temperature. He found that they were not due to oxidization of Fe^{+2} to Fe^{+3}, but stated that they may be attributed to alterations of the grain boundaries.

Conductivity mechanism

It is possible to argue about the conductivity mechanism at temperatures above 500–600°C. Denoting A_1, σ_{01} as the low, and A_2, σ_{02} as the high temperature values, respectively, it can easily be found from Table 1 that $A_2/A_1 \sim 2$–3, and $\sigma_{02}/\sigma_{01} \sim 10^7$–$10^9$, both values being typical for impurity ionic—intrinsic ionic conduction processes (Lidiard, 1957; Ioffe, 1960). If one considers the high-temperature interval as the region of an intrinsic conduction, σ values in it should be strictly reproducible. That this was not found in the experiments may be due to differences in composition of the samples and/or errors arising from dimension and resistance determinations. Despite the differences, scatter in the data appears to vary systematically with temperature, the errors being least at high T, which was confirmed by measurements repeated on different cuts of the same piece of rock.

However, as no measurement of the conductivity mechanism was performed, it is not possible to tell to what degree the above interpretation is correct, or whether another mechanism plays a role in the high-temperature region, too.

Effects of hydration and dehydration

The striking dissimilarity in character of graphs of $\log\sigma$ versus $1/T$ of samples with a high percentage of serpentinized olivine in the temperature range up to 545–625°C from those of unaltered rocks suggests that conductivity within this temperature interval is strongly influenced by the serpentine content. In the course of experiment no measurement was made to examine this phenomenon. However, previous measurements on serpentinites taken under ordinary pressure (Noritomi, 1961; Dvořák, 1967; Dvořák and Parkhomenko, 1971), have shown that the electrical conductivity of serpentinites is characterized by high values throughout the whole temperature interval from about 200°C up to 530–775°C with variations usually smaller than half an order of magnitude. This interval is followed by a sharp decrease of σ, which continues until the conductivity lies in the range given by A between 0.6 and 1.7 eV, and $\log\sigma_0$ between -2.4 and $+2.3$; both ranges are in agreement with those found for olivine-bearing rocks (Lubimova and Feldman, 1970). X-ray studies and differential thermal analyses (Kiyoura and Ito, 1954; Brindley and Zussman, 1957; Kobayashi, 1962), showed that the serpentine minerals of the overall composition $Mg_3Si_2O_5(OH)_4$, in the temperature interval 500–700°C (compare with 530–775°C found in the electrical conductivity experiments), exhibit an endothermic peak due to the loss of constitutional water. Since forsterite is the only resulting crystalline phase observed, this reaction is likely expressed by

$$Mg_6Si_4O_{10}(OH)_8 \rightarrow 3\ Mg_2SiO_4 + 4\ H_2O + SiO_2.$$

Probably some MgO is associated with SiO_2 (Brindley, 1963), and at higher temperatures between 790 and 850°C, enstatite can be formed. Serpentine minerals have a layered structure in which each layer is composed of one sheet of linked SiO_4 tetrahedra and one sheet of MgO(OH) octahedra, while forsterite consists of discrete SiO_4 tetrahedral groups joined by MgO octahedral groups. During the transformation, the MgO(OH) part of the serpentine layer must then undergo a considerable reorganization, which is accompanied by a collapse of the serpentine layer structure to the three-dimensional forsterite structure (Brindley and Zussman, 1957).

On the basis of the above evidence, we conclude that the rather irregular variation of conductivity with temperature in highly serpentinized samples is due to the high conductivity of serpentine and dehydration of this mineral between 500 and 700°C, which results in formation of forsterite. The decrease of σ in this temperature interval, then, should be regarded as the result of the structural differences between the starting serpentine and the end product olivine (forsterite). Further variations of σ with temperature are determined mainly by the resultant olivine. Comparison of the high-temperature characteristics reveals close agreement between those for olivinites and peridotites, and those for dunites (Table 3). The overall shift in the plots of $\log\sigma$ versus $1/T$ with pressure toward lower temperatures probably results also in lowering the beginning of the serpentine—forsterite transformation to 470–560°C at 20 kb

Table 3. Comparison of the high-temperature characteristics of dunites and those of olivinites and peridotites

	$\Delta A/\Delta P$(eV kb^{-1})	$\Delta\log\sigma_0/\Delta P$(kb^{-1})	$\Delta(\Delta\log\sigma/\Delta P)/\Delta T$(kb^{-1} °K^{-1})
Dunites	-5.0×10^{-3}	-22.0×10^{-3}	-3.9×10^{-5}
Olivinites and Peridotites	-5.37×10^{-3}	-21.8×10^{-3}	-3.6×10^{-5}

and, perhaps, in narrowing of the width of the reaction interval.

It is possible to imagine a tectonic process which will bring the olivine-rich mantle material, such as peridotite or dunite, into the crust. It is also possible to suppose that if this material is under partial water pressure, the hydration of olivine will occur. As the temperature increases with depth, an isotherm of about 500°C (or less, due to possible lowering of the reaction temperature with pressure) is reached which marks the beginning of dehydration, and according to the proposed idea, a decrease of conductivity. After the dehydration is completed, conductivity roughly follows that of olivine. There is also a possibility of formation of enstatite (Brindley, 1963) at higher temperatures which would probably lead to an increase of conductivity because of higher values of σ for enstatite than for olivine, as suggested by results obtained at low temperatures (up to about 250°C) (Dvořák and Schloessin, 1972).

Temperatures inferred from in-situ conductivity measurements

The shift of $\log\sigma$ versus $1/T$ graphs toward low temperatures can be of importance for any problem dealing with the depth variation of electrical conductivity and temperature in the uppermost parts of the earth, such as interpretation of data obtained by magnetotelluric and geomagnetic variation methods. In conductivity models derived from these studies, a value of 10^{-3} ohm^{-1} cm^{-1} (10^{-12} emu) has, for reasons which will not be discussed here, usually received special attention. It has been recognized that such a conductivity corresponds to a depth of the 1200°C isotherm (Uyeda and Rikitake, 1970). Also, the pressure effect has been assumed to be sufficiently small as to be dirsegarded. This assumption is not justified, as follows from Figure 5, which was constructed using measured conductivities assuming the extrapolation of σ values to higher temperatures as permissible. The temperature at which the conductivity of 10^{-3} ohm^{-1} cm^{-1} is reached appears to decrease with increasing pressure, leading, thus, to much lower values of T at a given pressure (depth). Any value of σ can, of course, be treated in a similar way. Let us take, for example, the Alert anomaly which has been interpreted as a rise of highly conductive mantle material ($\sigma\sim10^{-2}$ ohm^{-1} cm^{-1}) to a level of 25–30 km below the surface, and consequently as an elevation of the 1400–1500°C isotherm to this particular level (Rikitake and Whitham, 1964). For an estimated pressure of about 8 kb at the depth of 25–30 km, our data give a temperature of about 760–1050°C.

Magnetotelluric measurements made by Hermance and Grillot (1970) in southwestern Iceland can serve as another example. The authors used regional enhancement of temperatures in the crust and upper mantle as a possible explanation of their results. For the electric interface at a depth of 15–20 km coinciding with the crustal basalt—upper mantle peridotite boundary, they obtained, from room pressure measurements, temperatures between 820 and 1120°C. Estimating the pressure at a depth of 15–20 km to be about 4 to 6 kb, and using Watanabe's data (1970) for basalt with partial water content, one finds the required conductivity of $5.0–6.7\times10^{-4}$ ohm^{-1}

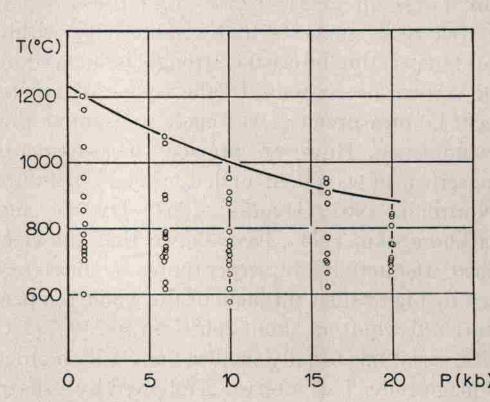

FIG. 5. Temperature corresponding to the conductivity of 10^{-3} ohm^{-1} cm^{-1} as a function of pressure.