PROBING THE HOLOCENE-WISCONSIN BOUNDARY IN POLAR ICE SHEETS

(Abstract)

by

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INTRODUCTION

Electrical resistivity measurements have been made at several locations on the Ross Ice Shelf and at Dome C, Antarctica. Measurements at stations on the Ross Ice Shelf (Bentley 1977, 1979, Shabtaie and Bentley 1979) and at Dome C (Shabtaie and others 1982, Shabtaie and Bentley 1983) have already been reported. A computer program using a linear filter method has been developed by S Shabtaie to calculate apparent resistivity on both floating and grounded ice sheets in which the actual resistivities, as well as other physical parameters, vary continuously with depth. In this paper we report on measurements made at Dome C and two other stations on the Ross Ice Shelf (R.I. and J9).

The field method, data analysis, and modeling techniques are similar to those described in the above references. In previous analyses the actual resistivity of the ice $\rho(z)$ was assumed to vary only with temperature, according to an Arrhenius law, and density, according to mixing laws for heterogeneous media. Analysis of resistivity sounding data from stations on the Ross Ice Shelf has now shown, however, that other physical and chemical changes in the ice could cause variations in resistivity with depth and must, therefore, be taken into account. Resistivity models based only on variations of temperature and density do not match the observations well. This was found to be the case particularly for stations located on ice streams and major outlet glaciers feeding the Ross Ice Shelf. In the current analysis we attempt to solve these problems by incorporating resistivity models that take into account the effect of physical and chemical changes on resistivity. At locations from which cores, whose physical and chemical properties have been measured, have come we have tried to correlate those properties with model resistivities.

To obtain more information about the dependence of resistivity on density and to examine further the highly resistive layer, electrical soundings were carried out at Dome C. Since the ice at Dome C is about 3 500 m thick, resistivity profiles with separations of many kilometers were needed to obtain information about the existence of highly-resistive ice layers and the bedrock beneath.

During the 1977-78 field season, drilling was completed down to a depth of 905 m at Dome C. Temperature and density-depth measurements are available for investigations of the dependence of resistivity on these factors. Furthermore, physical and chemical analyses of these cores were undertaken by many investigators. These analyses show pronounced differences between Wisconsin and Holocene ice.

RESISTIVITY MODELS

Dome C station (74°39'S 124°10'E, East Antarctica)

The measured apparent resistivities ρ_a are shown in Figure 1(a), along with calculated ρ_a models. Figure 1(b) shows the actual resistivities ρ for each model. Model 1 was calculated taking variations of density and temperature only into account. Temperatures were calculated from a steady-state model with bottom temperature at the pressure-melting point, using an activation energy of 0.25 eV. The other models in Figure 1 use the same temperature and activation energy as model 1, but also include additional changes in ρ with depth in the ice that have been chosen to provide better fits to the observed data, and at the same time to accord in some way with other measured physical or chemical properties at Dome C. This requires further explanation.

Our procedure was to examine first the variation of impurities with depth in the core. However, no model of $\rho(z)$ generated using resistivities inversely dependent on the concentration of impurities such as salts (Petit and others 1981), sulfates and acids (M Legrand personal communication), microparticles (Mosley-Thompson and Thompson 1980), and CO_2 (Delmas and others 1980) could produce ρ_a models that match the observed data, because those impurities all show an increase with depth in the Wisconsin ice. On the other hand, there would be no physical basis for assuming a direct correlation between conductive impurities and $\rho(z)$.

We therefore looked next at the measured physical properties. The principal change is in the crystal size (Duval and Lorius 1980). Crystals increase in size between 90 and 400 m, and below 540 m; in between there is a marked decrease. The abrupt decrease in crystal size between 400 and 540 m is associated with the Holocene-Wisconsin boundary as determined by the oxygen isotope ratios (Lorius and others 1979). In this case, we found that the observed apparent resistivities could be fitted with models of $\rho(z)$ that were qualitatively similar, i.e. that showed the same saw-tooth shape as the crystal-size variation in their deviation from the basic $\rho(z)$ model (model 1). This is shown in Figure 1 by both model 2 and model 3, which do not differ at depths or separations less than 1 500 m.

The lower curve, marked model 2 in Figure 1(a), clearly does not fit the observed apparent resistivities at separations greater than 2 or 3 km. To produce a fit, it is necessary to increase the resistivity somewhere at considerable depth, either in the ice (model 3 below a depth of 1 700 m) or in the underlying bedrock. The upper curve, marked model 2 in Figure 1(a), is based on the same resistivities in

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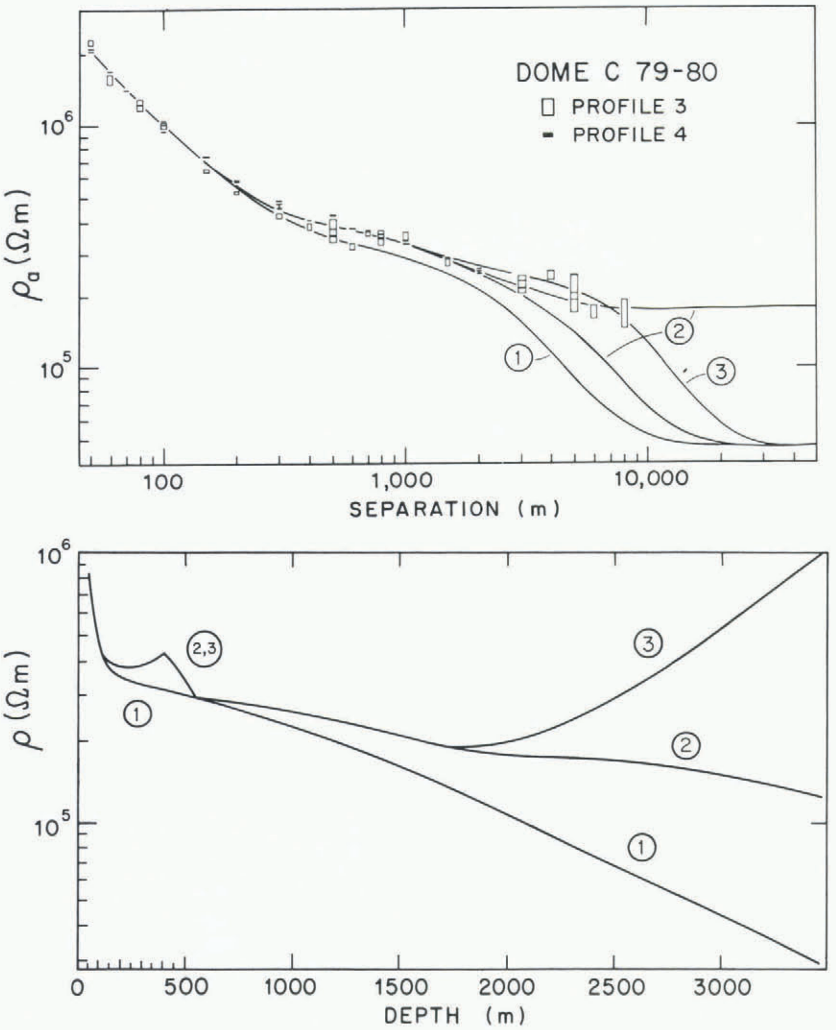

Fig.1. (a). Plot of apparent resistivities vs electrode separation for two profiles at Dome C. The height of each symbol represents the standard deviation calculated from linear fits to plots of potential difference vs current. Only data at separations greater than 50 m are shown. The solid curves show apparent resistivities calculated from the actual resistivity-depth models shown in Figure 1(b). The upper and lower branches of model 2 correspond to assumed resistivities in the bedrock of 200 and 50 kΩ m, respectively. (b). Actual resistivity vs depth in the ice sheet according to three different models (explained in the text).

the ice as the lower (model 2) curve, but includes a very high resistivity of 200 kΩ m in the bedrock. Both the lower branch of model 2 and model 3 incorporate a bedrock resistivity of 50 kΩ m, still a high value (Shabtaie and others 1982). Resistivities lower than 50 kΩ m in the bedrock can be accommodated by assuming still higher resistivites in the deep ice.

Although there is no obvious reason why resistivity should increase with crystal size, it is possible that variation may have been controlled by climatic change, i.e. that the history and age of polar ice influences its electrical properties. Regardless of the cause(s) of the resistivity changes, the correlation suggests that resistivity measurements could be used to estimate the depth to the climatic boundary.

Station J9 (82°21'S 168°39'W, Ross Ice Shelf, Antarctica)

Station J9 is the site of the Ross Ice Shelf (RISP) drill camp and is downstream from ice stream B. Limited information on isotopic ratios ($\delta^{18}O$) (P M Grootes personal communication), crystal cross-sections (I Zotikov and A J Gow personal communication), and ionic impurities (Herron and Langway 1979) are available. Both the isotopic ratios and the crystal cross-sections reveal a distinct climatic boundary at about 275 m depth.

The measured ρ_a profile and four different models are shown in Figure 2(a). All models are based on temperature and density measured from the bore hole, with the activation energy of 0.25 eV for both snow and solid ice. The calculated ρ models are shown in Figure 2(b). Model 1 is calculated according to the temperature and density variations alone. The resulting ρ_a model is lower than the measured values at separations greater than 200 m. Next we considered the variation of salt impurities with depth reported by Herron and Langway (1979). ρ is assumed to vary with concentration of impurity according to the relationship obtained by Gross and others (1978) and discussed by Shabtaie and Bentley (1979). The resulting model 2 does not match the measured values for separations greater than 50 m, so it seems unlikely that the reduction of impurities with depth in the J9 hole is responsible for the increase in resistivity.

Models 3 and 4 both fit the observed data. For model 3, the resistivity is taken to vary with depth in accordance with changes in crystal cross-sections observed from the core hole. Model 4 shows a resistivity increase that takes place instead entirely in the Wisconsin-age ice. There are no impurity measurements at this depth to confirm or rule out this model, but we suspect that, as at Dome C, Byrd, and Camp Century stations, the impurities in the J9 core increase in the Wisconsin, causing a reduction rather than an increase in resistivity. Therefore, model 3 is preferred. Unfortunately, the strong effect on the apparent resistivities at large electrode separations of the highly-conductive sea-water beneath the relatively thin ice shelf makes it difficult to pinpoint transitions deep within the ice.

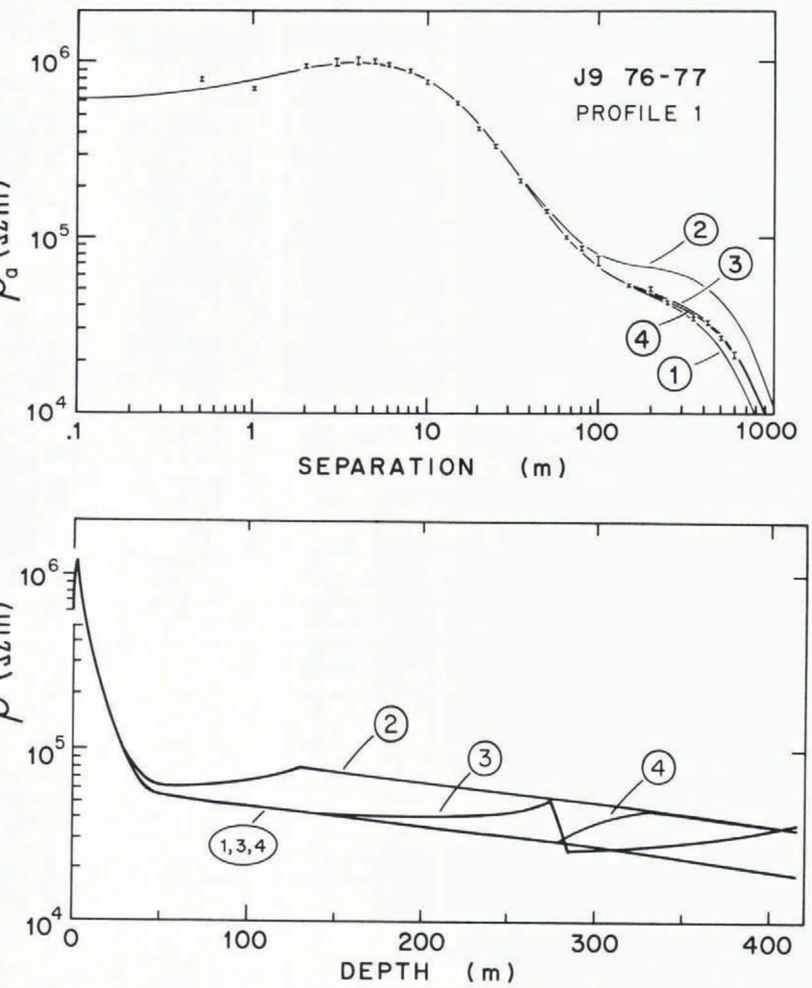

Fig.2. Plots of (a) apparent resistivity vs electrode separation, and (b) actual resistivity vs depth at station J9. Solid curves are calculated from models described in the text.

Station R.I. (80°11'S 161°35'W, Ross Ice Shelf, Antarctica)

This station is located downstream from either ice stream D or ice stream E (the flow bands are not well-defined). The ice is thicker here than at J9, so resistivity variations within the ice column are more easily seen. The results of electrical soundings on two profiles perpendicular to each other are shown in Figure 3(a). The ρ_a values were combined statistically (Shabtaie and Bentley 1979) beyond an electrode separation of 15 m where the effect of seasonal temperature changes and upper surface density inhomogeneities are absent. Five ρ and corresponding ρ_a models were chosen in this case (Figures 3(a) and (b)). As at the other stations, model 1 is based on density and temperature alone. (Temperatures were calculated assuming a steady-state ice shelf with zero bottom balance rate.) Model 2 is the same as model 1, except that a bottom melt rate of 0.5 m a^{-1} was used for the steady-state temperature calculation (this model could also be thought of as reflecting colder temperatures in a non-steady-state ice shelf. Clearly, the resistivities that result even from a large melt rate are not nearly high enough to match the measured data. Therefore, additional factors again must be taken into account.

Models 3, 4, and 5 all yield apparent resistivities that fit the data satisfactorily. All exhibit very high resistivities near the bottom of the ice shelf, as is required by the relatively high apparent resistivities at separations greater than 500 m. They differ substantially, however, in their characteristics at intermediate depths. Model 3 is the simplest, showing only a gradual monotonic increase in resistivity. This model satisfies the observations well. Model 4 was chosen to be similar to the variations fitted at Dome C and J9; it shows that that shape, with a Holocene/Wisconsin transition between 300 and 325 m deep, is also compatible with the apparent resistivities. Model 5 is a more extreme case, originally selected on a curve-fitting basis alone; it is included to give a sense of the range of models that can satisfy the observations. Because of its similarity to the models fit at Dome C and J9, where borehole data exist, we prefer model 4 for this site.

CONCLUSION

Measurements at several locations in Antarctica have resulted in the discovery of high-resistivity anomalies that could be correlated with the transition from Holocene to Wisconsin ice. The correlation is good at two locations (J9, on the Ross Ice Shelf, and Dome C, in East Antarctica), where the transition depth is known from studies of the physical and chemical properties of ice cores.

The effects of the variations of density and temperature with depth have been included in our analyses, but the calculations show that these two factors alone are not sufficient to explain the observations. Higher resistivities are required.

The cause of this resistivity increase is not well-known, and may be attributed to any of the several physical and chemical changes that occur in the ice sheet. The crystal size increases with depth down to the transition zone, decreases rather sharply over a short depth, and thereafter increases again. Resist-

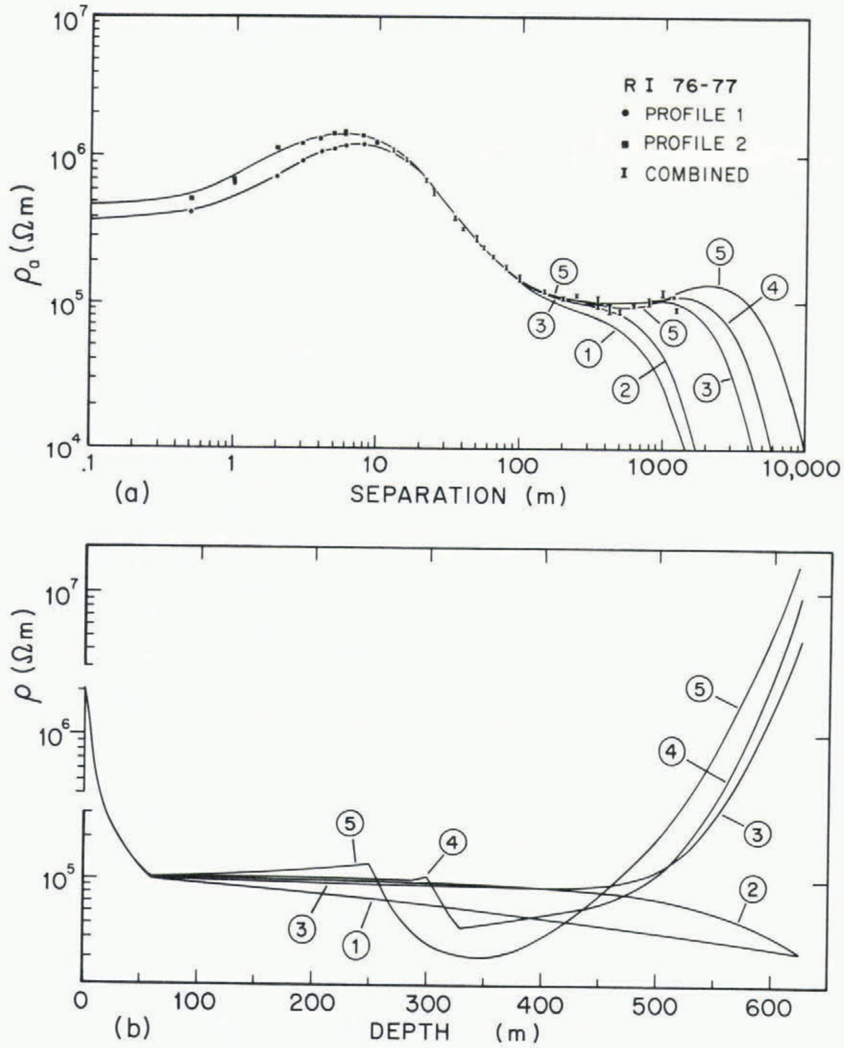

Fig.3. Plots of (a) apparent resistivity vs electrode separation, and (b) actual resistivity vs depth at station R.I. Solid curves are calculated from models described in the text.

ivities vs depth can be modeled similarly. No correlation was found between the changes in chemical impurities and the resistivity increase. If the resistivity variations are due to crystal-size changes, one might use the technique to obtain an age-depth curve since the crystal growth in an ice sheet is time-dependent.

Regardless of the cause(s) of the resistivity changes, the in situ observations can be used to estimate the depth at which this boundary exists. This technique is especially suitable for thick, grounded ice sheets where transition zones are well above the bedrock, and on ice shelves at locations close to inflowing ice streams and major outlet glaciers where the ice is relatively thick. The method can perhaps be used to trace the climatic boundary along flowlines.

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