Geomagnetic variation studies in the southern Appalachians: preliminary results

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Magnetic field variations have been recorded at 12 stations located on a traverse across the Blue Ridge and the Piedmont provinces of the southern Appalachian system in Georgia. The records were Fourier analyzed, and the transfer functions, relating the anomalous vertical magnetic field variations to the horizontal field, were obtained for different periods ranging between 16 and 4000 s. The transfer functions for the southeasternmost stations seem to be affected by the coast effect and possibly by the Coastal Plain sediment. In the northwestern part of the survey, the direction of the transfer functions was away from the coast. This suggests the presence of a strong conductor in the lower crust or the upper mantle beneath the Blue Ridge and extending farther toward the northwest.

Les variations du champ magnétique terrestre ont été enregistrées à 12 stations situées sur une ligne traversant les provinces du Piedmont et du Blue Ridge dans le sud des Appalaches, en Géorgie. Les enregistrements furent décomposés en fréquence et des vecteurs fonctions de transfert, reliant les variations anormales du champ magnétique vertical au champ horizontal, ont été déterminés à différentes périodes comprises entre 16 et 4000 s. Les fonctions de transfert des stations du sud-est semblent être affectées par un effet de côte et peut-être aussi par les sédiments des plaines de la côte. Dans le nord-ouest, les vecteurs de transfert pointent dans la direction opposée à la côte. Cet effet suggère la présence d'un conducteur puissant dans la croûte inférieure ou dans le manteau sous la province du Blue Ridge; la limite nord-ouest du conducteur n'a pas été détectée.


Introduction

The southern Appalachians consist of several distinct geologic belts trending from southwest to northeast (see Fig. 1). From the foreland area of the Cumberland Plateau in the northwest to the Coastal Plain in the southeast, the major provinces are: the sedimentary Valley and Ridge, the Blue Ridge, and the crystalline belts of the Piedmont. The Piedmont has been subdivided into the Inner Piedmont, the King’s Mountain, and the Charlotte and the Carolina Slate belts (e.g., King 1955; Rodgers 1970).

The Valley and Ridge Province consists mostly of folded and thrust faulted sedimentary rocks of Paleozoic age. The interpretation of the geology and of the seismic reflection surveys suggests that the deformations, in this province, are mostly thin skinned (Harris 1976; Harris and Milici 1977). The Blue Ridge is predominantly metamorphic, although some unmetamorphosed sediments are exposed locally, such as the Grandfather Mountain in Tennessee (Hurst 1973). The Brevard Zone, a linear band of cataclastic rocks extending from Alabama to Virginia, separates the Blue Ridge from the Piedmont (Hatcher 1975). The Piedmont Plateau, in which four major belts can be distinguished, is the crystalline core of the Appalachians (King 1977). The Inner Piedmont consists of high-grade metamorphics with some large plutons, such as the Stone Mountain and the Elberton granites; the King’s Mountain and its continuation, the Pine Mountain Belt, form a narrow band of volcanic and metasedimentary rocks; the Charlotte belt is made of metasedimentary rocks with some plutons; and the Carolina Slate belt consists predominantly of metavolcanics.

To the southeast of the Piedmont, the Atlantic Coastal Plain, in part submerged, forms a wedge of Cenozoic, mostly clastic sedimentary rocks, thickening toward the ocean. The average sediment thickness is 2–3 km but, in a few troughs, it may be as much as 6 km (Marine and Siple 1974). Wells drilled through the young sediment of the Coastal Plain indicate that the basement is made up of metasedimentary rocks of the same type as those exposed farther north (King 1975).

The transition between the continent and the ocean is not very well defined (Keen 1974). Between the continental shelf, which extends about 100 km into the Atlantic, and the Abyssal Plain lies the Blake Plateau. It is an elevated, 250 km wide region where 10 km of sediment overlies a basement apparently of oceanic composition (Sheridan 1974). The total crustal thickness beneath the plateau is about 25 km.

After Wilson (1966) suggested that the Atlantic Ocean closed and reopened, several attempts have been made to interpret the geology of the southern Appalachians.
The survey region (central square) is detailed in Fig. 2.

in terms of plate theory (Bird and Dewey 1970; Hatcher 1972, 1978; Dewey and Kidd 1974; Rankin 1975; and many others). No consensus has been reached yet on a detailed tectonic history of the Appalachians, but the use of geophysical techniques has provided new information on the structural setting of the southern Appalachians. A seismic refraction profile between Campbell, Tennessee and Dahlonega, Georgia (Warren 1968) shows that the crustal thickness increases from about 40 km in the Cumberland Plateau to more than 50 km in the Blue Ridge. The large negative Bouguer anomaly (Haworth et al. 1980) supports the suggestion that the crust is thicker than 50 km beneath the Blue Ridge. Recently, a detailed picture of the upper crust has been obtained by the seismic reflection study conducted by the Consortium for Continental Reflection Profiling (COCORP) (Cook et al. 1979). The traverse extends roughly between Madisonville, Tennessee and Augusta, Georgia but is not continuous, and there is a major data gap in the Valley and Ridge and in the northern part of the Blue Ridge. Additional seismic lines have been obtained more recently in the Georgia Coastal Plain (Cook et al. 1981). The seismic cross section (Cook et al. 1979) shows, beneath the Piedmont, southeasterly dipping reflectors that have been interpreted as buried sedimentary layers; the dip of the reflectors suggests that these sediments are the same as are exposed farther north in the Valley and Ridge and locally in the Blue Ridge, and that the major tectonic feature is an overthrust of the order of 250 km. There are different viewpoints on the nature of the reflectors detected by COCORP. Cook et al. (1979) suggested that these are relatively unmetamorphosed rocks. This hypothesis has been challenged by other researchers (e.g., Reed and Bryant 1980; Ellwood et al. 1980). They argued that the structural and metamorphic relationships indicate that these reflectors are metasedimentary rocks, whose grade is at least greenschist. The major feature of the magnetic anomaly map of Zietz et al. (1980) is a remarkably straight offset of the magnetic contours that extends from Alabama to New York (Zietz and King 1977). This lineament cuts across geological boundaries; in Alabama, Georgia, and Tennessee it is part of the Valley and Ridge Province.

Additional constraints on the nature of the crust in the Appalachians are provided by different geophysical methods. A correlation between lower crustal electrical
conductivity anomalies and past tectonic activity has been suggested by Law and Riddihough (1971). This hypothesis has been generally accepted (e.g., Gough 1973; Garland 1975, 1981; Edwards et al. 1981), but the cause of the conductivity anomalies, in regions that are no longer active, is still not clear; it has been proposed that they are related to, among other things, highly metamorphosed rocks or subducted oceanic crust.

An electrical conductor has been detected in the lower crust of southern Scotland by geomagnetic (Bailey and Edwards 1976) and by magnetotelluric studies (Jones and Hutton 1979). These authors suggested that this anomaly is related to the existence of remnants of the proto-Atlantic ocean, Iapetus (Dewey 1974). In eastern North America, there is good evidence for a highly conductive lower crust in the northern Appalachians, in Canada (Cochrane and Hyndman 1974; Bailey et al. 1974, 1978), and in the central Appalachians in Virginia (Edwards and Greenhouse 1975). Greenhouse and Bailey (1981) reviewed the available geomagnetic variation studies in the eastern United States. They concluded that there is a major crustal conductivity anomaly extending from the New York — Vermont state line southwestward into Pennsylvania and Virginia. The eastern edge of the anomaly corresponds more or less to the boundary between the crystalline and sedimentary Appalachians. Several models of the crustal conductivity indicate that the source of the anomaly is in the lower crust, where the resistivity could be as low as 2 ohm·m. Until recently no geomagnetic variation data from the southeastern United States were available, and the southern extension of the conductivity anomaly described by Greenhouse and Bailey (1981) was not defined.

Recently Thompson et al. (1983) conducted a controlled-source study of the eletrical conductivity of the upper crust around the Charlotte belt and the Carolina Slate belt in Georgia. They found that the electrical conductivity increases at a depth of 10—15 km; they propose that this higher conductivity is related to the buried sediment detected by the COCORP survey of the Georgia Piedmont.

The purpose of this paper is to report on a geomagnetic study that has been carried out during the past years in northeastern Georgia. During this survey, the variations of the natural electric and magnetic fields have been recorded at 12 different sites. The main objective of the study was to detect a possible southwestward extension of the conductor described by Greenhouse and Bailey (1981). The vector transfer functions, determined at different periods ranging between 15.2 and 3867.4 s, indicate a high conductivity in the lower crust or the upper mantle beneath the Blue Ridge. The northwestern edge of this conductor has not been determined, and it is possible that it extends beneath the Valley and Ridge.

The determination of the transfer functions

Following the work of Parkinson (1959) and Schmucker (1964), many studies have been conducted to determine the Earth’s conductivity structure from geomagnetic variations. The method consists of determining relationships between different components of the magnetic variations and to separate them into a “normal” and an “anomalous” part. A uniform magnetic variation at the surface of a horizontally layered conductive half space should have no vertical component. If geomagnetic field variations are measured at the Earth’s surface, a vertical component indicates that the geomagnetic variation is not uniform, that the conductivity varies laterally, or both. Systematic measurements in mid-latitudes have shown that the magnetic field variations are uniform over reasonably large regions (e.g., Schmucker 1964) and that, therefore, the anomalous vertical field is mostly of internal origin. For each frequency, the vector transfer function (A, B) at one site is defined as the horizontal component that is best correlated with the vertical field (e.g., Schmucker 1964; Everett and Hyndman 1967). It is a complex vector in the horizontal plane that is determined only by the conductivity structure of the region of induction, provided that the incident geomagnetic variation is uniform and that the field at each station is dominated by the external inducing field. As mentioned above, the first of these assumptions is generally valid in the mid-latitudes and the second is reasonable as long as the anomalous field is small: in the present study this latter assumption is supported by the small amplitude of the transfer functions (Hyndman and Cochrane 1971).

Three sets of three-component fluxgate magnetometers (E.D.A. FM 100) were used to measure the magnetic field. The telluric variations were also recorded at all the stations but one. The field measurements were digitized, multiplexed, and recorded on magnetic cassettes. The data were collected at 12 different sites in northeastern Georgia between Louisville and Young Harris. At each station, the data were recorded at a fast rate of 32 samples per minute for about 12 h and a slower rate of 1 sample per minute for 2—3 weeks. The station locations are displayed in Fig. 2 on a map where the boundaries between the major geologic provinces are delineated (the approximate location of the COCORP traverse is also shown on this map). Because the recording stations were always left unattended except for periodical checks, several problems affected the data collection. The most common difficulty was that the telluric lines were often cut by rodents or hunters, and there are many gaps in the telluric data. For this reason, the magnetotelluric data have not been completely analyzed yet and this report will discuss only the interpretation of the magnetic variation study.

After recording, the data were transferred onto a computer tape for processing and interpretation. The records were divided into shorter sections of 2048 samples, which were Fourier analyzed separately. From
FIG. 2. The map of the survey area; the political boundaries and the different geologic belts are delineated. The survey sites are numbered from 10 to 120. The lines indicate the approximate location of the COCORP traverse.

The Fourier spectra, complex transfer functions were determined for each event following a method similar to those proposed by Everett and Hyndman (1967) and Schmucker (1970).

The practical method to determine the transfer functions consists of assuming that there exists, at each frequency, a linear relationship of the form

$$Z = AX + BY + \delta$$

between the Fourier components of $X$, $Y$, $Z$, the north, east, and vertical down magnetic fields; $\delta$ is a residual, and $A$ and $B$ are the components of the transfer function. They are estimated by minimizing the power of the residual $(\delta, \delta^*)$. The calculations (Everett and Hyndman 1967) show that

$$A = \langle Z, X^* \rangle \langle Y, Y^* \rangle - \langle Z, Y^* \rangle \langle Y, X^* \rangle$$

$$B = \langle Z, Y^* \rangle \langle X, X^* \rangle - \langle Z, X^* \rangle \langle Y, X^* \rangle$$

where $\langle Z, X^* \rangle$ is the cross power spectrum of $Z$ and $X$; the asterisk indicates the complex conjugate. The implicit assumption is that $\langle \delta, X^* \rangle = 0$ can be tested by determining the coherency between the predicted and measured vertical fields, and the data that do not satisfy the test are rejected. The transfer functions from all the different sections were then stacked for each station following the analysis of Bailey et al. (1974). They are expressed as a pair of real and imaginary induction arrows; the usual convention (in the northern hemisphere) to reverse the direction of the arrows so that they will point toward the current concentrations (Edwards et al. 1971) has been followed. A summary of the data used to compute transfer functions for each station is given in Table 1.

The real and imaginary induction vectors for three different periods (241, 772, and 3867 s) are presented in Figs. 3, 4, and 5, respectively. Some induction arrows are missing on the map because the data did not satisfy the coherency tests. None of the results obtained were eliminated but some are obviously in error. This is the situation with the 241 s induction arrow obtained at station 110 (near Tallulah Falls, Georgia). The telluric variations at this station also show erratic behavior; we do not have a definite explanation for this (there is no pipeline nearby). These maps present a relatively consistent picture. In the southernmost part of the profile, the real induction arrows point west-southwest; they point west-northwest in the northernmost part of the traverse. This trend does not change much from period to period: the amplitude of the transfer functions is smaller at the shorter periods and the direction of the arrows is slightly less consistent, as would be expected.

**Interpretation**

Many geomagnetic measurements made near the continent-ocean transition have shown a coast effect, characterized by the induction arrows pointing toward the ocean. Parkinson and Jones (1979) reviewed the data and some of the hypotheses on the causes of this effect (induction in sea water, different conductivity structures beneath oceans and continents). The effect is more pronounced in tectonically inactive regions, such as Western Australia (Everett and Hyndman 1967) or eastern Canada (Bailey et al. 1974).
The Atlantic coastline is about 200 km southeast of the end of the profile, but the shelf extends 100 km from the coast, and the Blake Plateau forms a transition region between continent and ocean. The 500 m bathymetric line (east of the Blake Plateau) is 450 km away. The surface rocks of the Coastal Plain are young sedimentary rocks, presumably more conductive than the crystalline rocks exposed farther north; they will enhance slightly the coast effect. It was thus expected that a small coast effect (with a southeastward transfer function of 0.2) would be observed. In the southernmost part of the profile, the induction arrows point to the southwest, probably as a combination of the effects of the Coastal Plain sediment, the Atlantic Ocean, and perhaps also the Gulf of Mexico. In the northernmost part of the profile, the induction arrows point to the northwest and suggest the presence of a conductor in that direction.

One possible interpretation of the induction arrows...
shown in Figs. 3, 4, and 5 is that a conductor is located to the west of the profile. This hypothesis was not retained because all the major geologic trends of the region follow a two-dimensional northeast–southwest striking structure. The departure from two dimensionality, manifested by the angles between the induction arrows and the northwest–southeast direction, can be explained by current channeling as long as these angles do not exceed 45° (Bailey et al. 1974). Additional data need to be collected to verify the assumption that the conductivity structure is two dimensional and follows the geologic trends. It was introduced because, at this stage, it is the only possible way to reduce the ambiguity of the geomagnetic depth soundings and to attempt a more quantitative interpretation. This interpretation was performed by computing, by the finite difference method, the transfer functions for two-dimensional conductivity models.
In order to fit these models to the data, the profile and the induction vectors were projected on a northwest-southeast plane. The projected transfer functions give the real and imaginary parts of the ratio of the vertical to the horizontal magnetic field variations in the direction perpendicular to the strike of the structure. They represent the response to a magnetic storm polarized in the direction of the horizontal field. The transfer functions projected on the direction 135°S are shown in Fig. 6 for the three frequencies considered. The trend is the same at all three frequencies; the amplitude increases northward and this occurs farther north at the shorter periods. As mentioned before, the direction of the induction vectors (or the sign of the projected transfer functions) is away from the coast and the Coastal Plain sediment. Another conductor in the crust, north of the Coastal Plain, is therefore needed to explain the observed geomagnetic variations.

Since the vertical magnetic field is largest near the edge of a conductor and the COCORP reflectors extend beneath the Coastal Plain, south of the arrow reversals, it is doubtful that the geomagnetic variations can be explained by the higher conductivity of the buried sedimentary rocks presumably detected by COCORP. However, several attempts were made, with limited success, to fit the data with models locating the conductor in the buried sediment, at a depth of 12 km. Several models with a uniform conductivity throughout the buried sediment were unsuccessful at predicting the observed transfer functions. A progressive thinning of the sedimentary layers toward the south did not explain the observations either. Lateral changes in the buried sediment conductivity provided a slightly better fit to the data. A resistivity lower than 20 Ω·m is necessary to fit the long-period data, but the transfer functions computed for the shorter periods are much larger than observed (i.e., a transfer function of 0.6 is predicted and the observed transfer functions vary between 0.2 and 0.3).

This change in sediment conductivity takes place in the Blue Ridge Province or farther north, but the data from the geomagnetic depth soundings do not require that the changes correlate with the boundary of a geological province.

Although the data do not rule out a lateral change in conductivity in the layers presumed to be buried sedimentary rocks, the authors found that a conductivity change in the deeper crust or in the upper mantle fits the data better. The resolution of the method is not sufficient to accurately locate the conductor and determine its conductivity, thickness, and depth. The successful models require the conductivity, in the lower crust or the upper mantle, to increase by two orders of magnitude beneath the Blue Ridge.

Two examples of satisfactory models are shown in Fig. 7. All the parameters of the models, with the exception of the deep conductor, were introduced because of geological or other geophysical constraints. Figure 7a shows a model with a laterally varying conductor in the lower crust and the results of this model compared with the observed data (the lower crustal conductivity increases from $3 \times 10^{-4}$ to $5 \times 10^{-2}$ S/m). Figure 7b shows the results for a model with a conductor in the upper mantle beneath the Blue Ridge where the conductivity increases from $10^{-2}$ to 0.5 S/m.

No difference in conductivity between the oceanic and the continental regions, other than the sea water, has been introduced in the model. The effect of a more conductive oceanic crust and mantle would need to be offset by an even stronger conductor. Little information is available on the resistivity of the Coastal Plain sediment, but the value assumed is within the range for young sedimentary rocks and does not strongly affect the conclusions. The resistivity of buried older sediment is likely to be higher than 100 Ω·m; an even stronger conductor below is needed to explain the geomagnetic variations. For the shorter periods, the transfer functions computed for the lower crustal conductor of Fig. 7a are larger than observed and the data are better fitted by the deeper conductor (i.e., in the mantle) shown in Fig. 7b. The fit of the model to the data was made on the real part of the transfer functions; the quadrature components are small and they could not be fitted by the models.

Conclusions and discussion

The geomagnetic depth soundings are poor at determining depths and thicknesses of buried conductors (Edwards et al. 1981); they provide better constraints on the lateral location. The two models in Fig. 7a and b illustrate this point quite clearly. Although in both models the horizontal location of the conductor is identical, the vertical distribution of the conductivity is radically different; therefore, the tectonic implications are dissimilar. Confinement of the conductor to the crust or to the mantle can only be done on the basis of independent evidence.

We prefer a crustal location for the conductor for two reasons. First, the conductivities suggested by the mantle conductor model in Fig. 7b are so high as to suggest that partial melting would be required. This is implausible in a tectonically quiet region. Second, localized depth-sensitive studies using controlled sources (e.g., Nekut et al. 1977) show that thick conductive zones with conductivities as high as 0.1 S/m can occur in stable cratonic lower crust. Our preferred model is therefore similar to that of Fig. 7a. This model can still be varied significantly: removal of the buried sediments postulated by Cook et al. (1979) would not significantly affect the results; raising the lower boundary of the conductor to the seismically determined Mohorovičić discontinuity (about 35 km) based on the COCORP data (Cook et al. 1981) is possible if the conductivity is suitably increased by a factor of about two or three.
Fig. 6. The projection of the transfer functions on the direction N45°W, which is roughly perpendicular to the axis of the Appalachians for periods of (a) 3867 s, (b) 771 s, and (c) 241 s. The data from station 110 are believed to be erratic.
Regardless of its depth and conductivity, the anomaly may be part of the eastern North American conductive anomaly reviewed by Greenhouse and Bailey (1981); this anomalous zone extends more or less the entire length of the eastern side of the stable craton within the continental United States.

Although a definitive discussion of the cause of the high conductivity is not feasible, some of the possibilities will be reviewed.

If the conductor is located within the buried sedimentary layers, a conductivity of the order of 0.1 S/m is needed near the center of the anomaly. Such a high conductivity in sediments is commonly found in geopressure zones, where it is caused by phase changes in the shales and formation of montmorillonite. The very low resistivity (less than 1 Ω·m) reported locally in the Gulf Coast sediment may be caused by such a mechanism (Vozoff 1972). It is questionable whether a geopressed zone in the underthrust, deeply buried, and perhaps metamorphosed sediment explains the high conductivity all along the Appalachians.

Another hypothesis relates the conductor to the presence of graphites or other mineralizations in the crust. The graphite could be produced by metamorphism of carbon in the deeply buried sedimentary rocks. Carbonate rocks are common within the Appalachian sediment, and graphic schists, which may have been brought from the deep sedimentary layers, have been detected near the Brevard Zone. Alternatively, the graphites or the mineralizations could be within the crystalline basement beneath the buried sediment. The rocks, exposed at the surface to the northwest of the Blue Ridge, are part of the old basement that has been overthrust; they contain graphic schists and different mineral ores. It is possible that strong metamorphism in the basement of the underthrust crust produced graphites and other metallic conductors. Camfield and Gough (1977) have documented that graphic schists in the basement are the source of the Great Plains conductivity anomaly. Recent experimental work reported by Shankland and Duba (1982) shows that the presence of even moderate amounts of carbon in the crust or the mantle could increase significantly the conductivity of crystalline rocks.

Buried oceanic crust (Drury and Nibblet 1980) has also been invoked to explain the electrical conductivity anomalies. The oceanic layer consists of water-saturated sediment, from ancient rift zones or continental margins, overlying the crystalline oceanic crust. The high conductivity is caused by water trapped in the sediment and (or) by the serpentinization of the crystalline rocks. If the electrical conductivity anomalies were caused by the buried oceanic crust, their occurrence throughout the Appalachians would be comprehensible, but the mechanism that emplaced pieces of the oceanic crust in the lower crust or upper mantle still needs clarification; it

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**FIG. 7.** Two of the models that fit the data and the observed and predicted transfer functions. The models include the structures that are known from geology and geophysics and will influence the observations. The resistivities shown represent: the Atlantic Ocean (0.25 Ω·m); the Coastal Plain sediments (surficial 100 Ω·m layer); the crystalline crust (3000 Ω·m); the buried sedimentary layers suggested by COCORP (100 Ω·m); and the upper mantle (1000 Ω·m). The 20 and 10 Ω·m structures in (a) and the 2 Ω·m structure in (b) represent the deep anomalous conductor discussed in the text.
would involve northwesterly dipping subduction during parts of the Appalachian orogeny.

The observations and their interpretation do not escape the common problem of the geomagnetic depth soundings; the nature of the electrical conductivity anomalies is extremely elusive and their significance far from obvious. However, they provide information about physical parameters not always sensed by other geophysical techniques. In the southern Appalachians, they are an intriguing feature and deserve more study. Additional soundings are currently being conducted in order to extend the traverse toward the northwest and determine the lateral extent of the conductor. Magnetotelluric data are being collected and their interpretation may resolve some of the depth ambiguity. Another traverse should be conducted to the west of this one to check the validity of the two-dimensional hypothesis or to determine if a conductor is present to the west of the region. This additional information will increase our understanding of the conductivity anomalies detected by this and other geomagnetic surveys in the Appalachians.

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