occur in spring and late fall. Changes of water temperature at the intake during the months of summer stratification are often wind-related. Offstore winds (westerly or northwesterly) can cause upwelling which brings colder water to the intake such as on 6 and 11 September. Easterly or northeasterly winds during the months of stratification push warm surface water into the Duluth water intake area, causing higher water temperatures such as on 24 July.

A historical record of the types of amphibole minerals previously suspended in Lake Superior water may be derived from a study of the bottom sediments. Dell (*16*) reported hornblende as the predominant amphibole in the sand fraction of Lake Superior postglacial sediments with a trace of tremolite-actinolite also present in some cases. Our study (*17*) of the surficial sediments of western Lake Superior shows a clear pattern of a recently deposited sediment layer rich in cummingtonite-grunerite on top of older sediment which does not contain detectable amounts (< 1 percent) of cummingtonite-grunerite. This layer rich in cummingtonite-grunerite is thickest (90 m or more) and coarsest in the vicinity of a large taconite tailings discharge at Silver Bay, Minnesota (*18*). It spreads throughout much of western Lake Superior, becoming thin and diluted with other sediment at Duluth, which is at the western tip of the lake. Indication of recent changes in the mineralogy of suspended solids in western Lake Superior water is provided by our x-ray diffraction analysis of suspended sediment samples collected for several periods in the past by the City of Duluth water utility. Samples from 1939–1940 and 1949–1950 contain only trace amounts of amphibole with no detectable cummingtonite-grunerite, but all samples studied for the period 1964–1965 contained large amounts of amphibole (average, 31 percent of the total inorganic solids), most of which was cummingtonite-grunerite. The geological and limnological data indicate that the source of this large increase in amphibole material is the taconite tailings (*18*) that, since 1956, have been discharged into western Lake Superior at Silver Bay.

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14. The fiber counts were carried out by the Ontario Research Foundation (ORF), Sheridan Park, Ontario. A comparison with literature values is normally not possible since different preparation and counting methods are often used. Because ORF results have been reported for other water samples (*7*), these values can be compared with those results. The ORF fiber-counting technique consists of filtering the water sample with a 0.1-μm membrane filter, ashing the filter by maintaining it at 450°C for 3 hours, dispersing the ashed sample in 4 ml of distilled water, and centrifuging a 1-ml aliquot onto a carbon-coated electron microscope grid which is examined at $\times$ 25,000 magnification on a transmission electron microscope (Jeolco model JEM 100U) at 80 kv. The Environmental Protection Agency is currently developing a standard method for the counting of asbestos fibers in environmental samples.
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18. Cummingtonite-grunerite is found almost exclusively in metamorphic rocks, usually in metamorphosed iron formations. The eastern Biwabik iron formation in northeastern Minnesota has been contact-metamorphosed by the Duluth gabbro. B. M. French [*Minn. Geol. Surv. Bull.* **45**, 1 (1968)] has described in detail the formation of cummingtonite-grunerite in the metamorphosed iron formation near the Duluth gabbro. This cummingtonite-grunerite in many cases is acicular to asbestiform in habit and varies from iron-rich grunerite to magnesium-rich cummingtonite. (The infrared spectrum of a sample of cummingtonite-grunerite from the taconite iron ore body is identical to that of amosite asbestos. The infrared interpretation technique described by R. G. Burns and R. G. J. Strens [*Science* **153**, 890 (1966)] for cummingtonite-grunerite indicates that both samples have an Fe/(Fe + Mg) atom ratio of 0.76.) The taconite iron ore body has been mined, and, after the extraction of magnetite, the amphibole-rich tailings have been discharged since 1956 into western Lake Superior at Silver Bay, Minnesota.

25 January 1974; revised 31 May 1974

Electrical Resistivity Variations Associated with Earthquakes on the San Andreas Fault

Abstract. *A 24 percent precursory change in apparent electrical resistivity was observed before a magnitude 3.9 earthquake of strike-slip nature on the San Andreas fault in central California. The experimental configuration and numerical calculations suggest that the change is associated with a volume at depth rather than some near-surface phenomenon. The character and duration of the precursor period agree well with those of other earthquake studies and support a dilatant earthquake mechanism model.*

Laboratory studies (*1*) have shown significant variations in electrical resistivity of rock samples subjeced to compressive stress. Fieldwork by Barsukov (*2*) showed that variations of up to 20 percent in apparent resistivity are associated with thrust-type earthquakes in the Garm region of the Soviet Union. Two years ago we initiated a study to monitor the deep resistivity across a strike-slip section of the San Andreas fault south of Hollister, California (Fig. 1). Frequent microearthquakes occur in this areas with hypocenters from 2 to 12 km deep.

To monitor resistivity changes to such depths requires surface electrode arrays of comparable dimensions. The field problems of current cable installation over such distances dictated the choice of the conventional dipole-dipole array (3). The current dipole, or transmitter (TMTR), was emplaced on the eastern side of the fault and potential electrodes were placed at receiver (RCVR) sites near the current dipole (RCVR 3) and at two locations 10 km (RCVR 2) and 15 km (RCVR 1) southwest of the transmitter. These locations are shown in Fig. 1.

The response of an array is function of any resistivity changes in the region encompassed by the array. In order to interpret these responses, we made numerical calculations of apparent resistivity over hypothetical two-dimensional models. These models were developed on the basis of a resistivity survey of the area and available geological information. The survey showed a pronounced contrast between the resistivities of granites west of the fault (500 ohm meters) and those of sedimentary and metamorphic rocks east of the fault (5 to 50 ohm meters). The numerical calculations indicated that by monitoring shallow resistivities such as those at RCVR 3, we could evaluate the relative contribution of any near-surface changes at the large arrays.

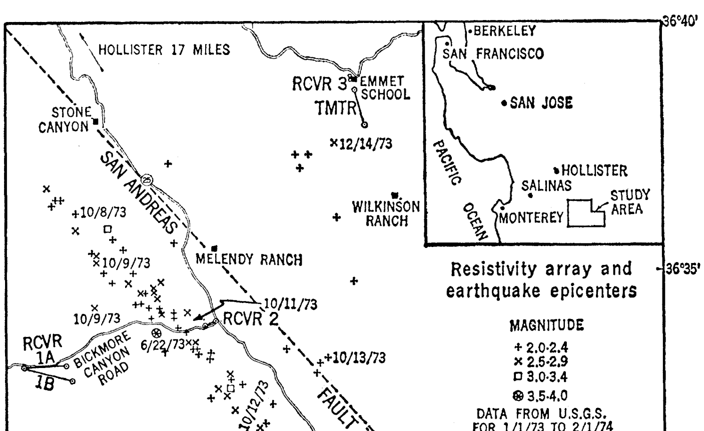
The current supply consists of an 85-kw motor generator set and a switching rectifier bridge producing a commutated d-c output. The current electrodes are sheets of aluminum (1 by 2 m) placed in pits, covered with soil, and soaked with salt water. The electrodes are 1.5 km apart and are connected to the rectifying bridge by copper cable (No. 4, American Wire Gage). A square-wave current with a peak to peak amplitude of up to 200 amperes and a period of 10 seconds is used. The difference in potential at the distant receiver sites is amplified and displayed on a chart recorder. During periods of low telluric current noise, signal amplitudes are averaged over many cycles. The current wave form is also monitored on a chart recorder, and both current and voltage measurements are calibrated with a precision reference voltage. Since the geometry of the array was not modified during the experiment, the most important information is the ratio of voltage to current. The data are normalized to the value obtained on 13 July 1973 and are plotted as normalized apparent resistivities in Fig. 2.

Telephone line telemetry was installed between the receiver and transmitter sites to improve the accuracy of the data. The signal voltage was telemetered to the transmitter site and synchronously detected and integrated. The same technique was applied simultaneously to the current wave form. This system improved the measurement accuracy to better than 1 percent during the periods when it was in service.

The normalized apparent resistivity values at receivers 1, 2, and 3 are plotted in Fig. 2 as a function of time from January 1973 to February 1974. Data from two methods of measurement are plotted. Circles represent chart recorder data comprising manually computed averages of up to 200 cycles. The error bars are plus and minus 2 standard errors. Triangles represent values obtained with the telemetered data synchronously detected and averaged for up to 450 cycles (1 hour and 15 minutes); the standard error of these values is less than 1 percent. Note that the triangles lie within the error bars of the data points calculated from the chart records.

Earthquakes of magnitude greater than 2.0 occurring during this time interval are located in Fig. 1 and are plotted in the lower portion of Fig. 2 (4). A recent analysis (5) has indicated that the epicenters should be relocated 2 to 3 km to the east, which places them on the surface expression of the fault.

Before the magnitude 3.9 earthquake on 22 June 1973 (the only earthquake



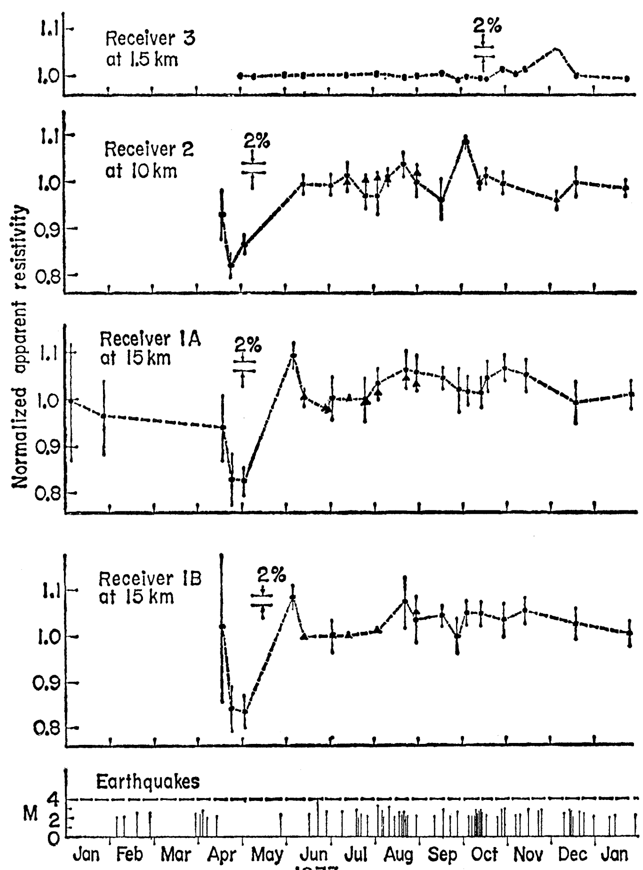


Fig. 1 (left). Location of current transmitter (TMTR) and three receiver dipoles (RCVR) used for monitoring deep resistivity across the San Andreas fault. Epicenters of earthquakes occurring during the experiment are plotted; dates refer to specific events discussed in the text. A recent analysis (5) has indicated that the epicenters should be relocated 2 to 3 km to the east. This places them on the surface expression of the fault. Fig. 2 (right). Normalized apparent resistivity as a function of time from January 1973 to February 1974. The earthquakes wnose epicenters are shown in Fig. 1 are plotted in the lower portion of the figure. The error bars represent plus and minus 2 standard errors. Absolute apparent resistivity values for the various receivers are receiver 1A, 81 ohm meters; receiver 1B, 75 ohm meters; receiver 2, 8 ohm meters; and receiver 3, 10 ohm meters.

in the magnitude range 3.5 to 4 shown in Fig. 2) a large variation in resistivity was observed at receiver 1A. While the data taken before 1 June 1973 are sparse in time and have relatively large errors, it appears that the apparent resistivity began decreasing in the middle of April, dropped by 10 percent, subsequently increased at the end of May to a maximum of 10 to 15 percent above the initial level, and dropped again to the initial level by 15 June. On 22 June, at the end of this 60-day variation, a magnitude 3.9 earthquake occurred midway between the transmitter and receiver at a depth of 9.5 km. An independent set of data from receiver 1B displays the same characteristics. The data from receiver 2 also show a pronounced variation but, possibly because of poor sampling, a maximum is not observed immediately before 22 June. The near-surface resistivity monitored at receiver 3 from 1 May shows no variation, which rules out the possibility that some near-surface phenomenon on the east side of the fault near the transmitter caused the variations at receivers 1 and 2. During the period 14 April to mid-October no rain fell in the area.

During the period 15 April to 15 June there was a reduction in the number of earthquakes in the area encompassed by the array. After the magnitude 3.9 event on 22 June, a more or less regular frequency of earthquakes was observed. This may be significant, since on the basis of the dilatancy model one might expect a dilatancy hardening effect and consequently a reduction of earthquake activity during the precursor period (*6*, *7*).

A second, less significant variation occurred at receiver 2 between 15 September and 15 October. A similar response was not observed at receiver 1. It is possible that this variation is associated with a swarm of earthquakes occurring within the array between 8 and 13 October. These are identified in Fig. 1. The total strain of these events is equivalent to that of an earthquake of magnitude 3.5 (*8*). The hypocenters ranged from 2.9 to 9.8 km in depth.

Finally, a 6 percent increase in the resistivity in the middle of December at receiver 3 may be associated with a magnitude 2.6 earthquake on 14 December. The epicenter of this event lay about 1 km southwest of the southernmost current electrode of the transmitter. This association, however, is clouded by the fact that the event occurred at a depth of 13 km, while the effective sensing depth of receiver 3 is of the order of 1 to 2 km. Considerable rain fell in the area starting in the middle of October. Very little change occurred at receiver 3 during this period, except just before the 14 December event. Thus, while the earthquake occurred at considerable depth, some phenomena associated with it may have occurred much closer to the surface.

The resistivity arrays used in this study allow only a partial definition of the locations and volumes of the sources that produce the observed variations in apparent resistivity. Numerical calculations indicate that either a large volume of earth was affected before the magnitude 3.9 earthquake, or the intrinsic resistivity of a smaller volume changed dramatically. For example, the results could be accounted for by an 80 percent change in intrinsic resistivity in a region 4 km wide and 2 to 6 km deep in the fault zone. Such a change is physically possible. The resistivity changes reflect changes in the porosity and degree of water saturation of the volume of rock stressed before an earthquake and consequently support the dilatational theory of earthquake mechanisms (*6*). The 60-day precursor time before the magnitude 3.9 event fits the relation between precursor time and magnitude given by Scholz *et al.* (*6*).

The sequence of an increase followed by a decrease in resistivity with a subsequent earthquake is similar to the response reported by Barsukov (*2*) for thrust faults in the Garm region of the Soviet Union (*6*). This suggests that a similar mechanism may be involved in the two different tectonic areas. However, a detailed interpretation of the observed phenomenon in terms of a dilatational or any other model should not be made until other events have been observed with greater accuracy and more sampling.

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Streamflow Greatly Reduced by Converting Deciduous Hardwood Stands to Pine

Abstract. *Fifteen years after two experimental watersheds in the southern Appalachians had been converted from a mature deciduous hardwood cover to white pine, annual streamflow was reduced about 20 centimeters (20 percent) below that expected for the hardwood cover. Streamflow was reduced during every month, with the largest monthly reductions (1.5 to 3.5 centimeters) occurring in the dormant and early growing seasons.*

Do different forest types vary substantially in evapotranspiration on a given site? This question has long been debated, and conclusive experimental evidence has generally been lacking (*1*). The results of experiments at the Coweeta Hydrologic Laboratory reported in 1968 (*2*) showed small reductions in streamflow and thus greater evapotranspiration only 10 years after mature hardwoods had been removed and eastern white pines (*Pinus strobus* L.) had been planted. The 5 years of additional data reported here provide conclusive evidence that evapotranspiration from the young white pine stands is substantially greater than from mature, deciduous forests. Evapotranspiration includes transpiration from plants, evaporation of water intercepted by plants and litter, and evaporation from soil.

The study site is in the southwestern corner of North Carolina in the Appalachian Mountains. Summers are cool, and winters are mild; rainfall is adequate in all seasons (*3*). The average annual precipitation varies from 2500 mm on the upper slopes to 1700 mm at the lower elevations. The two