Basic principles of electromagnetic and seismological investigation of shallow subduction zone structure

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Abstract. Much of the architecture of subduction zones consists of rock layers that are relatively uniform laterally, but not very thick. The identity of these layers is of primary interest in subduction zone studies because they bear on topics ranging from earthquake hazard assessment through whole-earth geochemical budgets to the causes of arc magmatism. Electromagnetic and seismological methods provide complementary ways to infer the properties of layers in the top few hundred kilometers of subduction zones because they are sensitive to different rock properties. Used in combination, they reduce the likelihood of overlooking significant structures in this region. The goal of this contribution is to introduce the underlying principles of the methods so that their strengths and weaknesses may be understood. Then some applications in subduction environments will be highlighted to illustrate these points.

The common feature of both electromagnetic and seismic methods is that they use waves to investigate the subsurface. The fundamental features of a wave are its speed $v$, its frequency $f$ and its wavelength $\lambda$, which are linked through the relation,

$$v = f\lambda.$$  \hfill (1)

The wave speed is a property of the material that the wave passes through. Electromagnetic waves travel through just about everything at a significant fraction of the speed of light, $3 \times 10^5 \text{ km sec}^{-1}$. In contrast, seismic waves travel at speeds between 5 and 8 km sec$^{-1}$. Thus, the above formula says that a wave oscillating at a frequency $f$ of 1 GHz ($1 \times 10^9 \text{ Hz}$) has a wavelength $\lambda$ of about 10 centimeters if it is a light wave, and about a micron ($10^{-6} \text{ m}$) if it is a seismic wave -- very different scales. Since a wave averages material properties over its wavelength, this means that the two methods of necessity probe the earth with very different frequency ranges and at different length scales.

Waves penetrate a substance until their energy gets absorbed and turned into heat. Electromagnetic and seismic waves also differ greatly in this characteristic. The former’s penetration ability is largely controlled by the electrical conductivity of the material. Highly conductive materials rapidly attenuate electromagnetic waves, but insulating ones do not as readily. With seismic waves, the attenuation depends more strongly in frequency. At frequencies lower than about 0.25 Hz, seismic waves easily travel through the whole earth. Above 5 Hz, a few hundred km is their limit. In consequence, we expect that it will be difficult to see through conductive regions of the crust to deeper ones with electromagnetic methods. It also will be hard to see features thinner than a few kilometers with seismic waves in the deeper parts of the Earth, because waves with short wavelengths attenuate rapidly on account of their high frequencies.

Of the various electromagnetic methods available with which to explore subduction zones, only the magnetotelluric (MT) method operates at low enough frequencies to sense the properties of the Earth at sufficient depth. For typical rock electrical properties, the frequencies of interest are $10^{-4}$-1 Hz. The Earth’s electric field oscillates at these
frequencies due to channelling of electrical disturbances between the ground and the ionosphere. The sources of the disturbances are lightning discharges, and electrical currents induced in the ionosphere by its interaction with the solar wind. What a MT measurement records is the relative strength of the electrical and the magnetic fields, and their phase lag, at a specific place, and at a specific frequency. The ratio of the field strength is turned into apparent resistivity of the ground $\rho_a$ and the phase lag into an angle $\phi$. On account for the phase difference between the magnetic and electric fields for insulators (in-phase, $\phi = 0^\circ$) and conductors (out-of-phase, $\phi = 90^\circ$), the phase information conveys a sense of whether the rock resistivity is increasing or decreasing with depth. The analysis task becomes, then, to find a one- two- or three-dimensional variation in rock resistivity properties that matches $\rho_a$ and $\phi$ measured along an MT profile.

MT studies provide two- or three-dimensional resistivity images. Material resistivities range over about 8 orders of magnitude, the low end of which corresponds to briny fluids, ore bodies and silicic melts, and the high end to dry crystalline rock. Consequently, conductive zones generally correspond to weak zones where fluids or melts are present in connected porosity or where fluids once were present and deposited interconnected conductive minerals (e.g. graphite, sulfides). Thus a straightforward way to interpret low resistivity patterns is as indicating zones of present or past weakness.

In contrast, seismological studies use the travel times of seismic waves emitted by artificial or natural sources to infer the wave speed structure of the Earth. Both P- and S-wave travel times can be used, but S-wave arrivals are usually harder to detect in artificial source studies and so are less frequently used in them. In addition to the travel times, valuable information lies in the shapes of the waves themselves. Their shapes are sensitive to the spatial changes in the seismic wave speed. Thus one can infer how abrupt a change in wave speed is, which is the principal way to detect lithologic boundaries. A useful rule of thumb is that the change in rock properties must happen in one-quarter of the wavelength $\lambda$ in order to be detectable at some frequency. In a seismological study, one combines the travel time and wave form information to build up a spatial map of wave speeds and interface positions and properties.

The traditional way to investigate a subduction zone margin is uses offshore controlled seismic sources and recorders located both on the seafloor bottom and onshore. This provides the seismic structure of the top few tens of kilometers of the subduction margin. Some of the onshore seismic instruments usually are used to record nearby or distant earthquakes to improve sensitivity to the deeper structure of the margin. This information comes from the seismic wave travel times, used to build up two- and three-dimensional images of wave speed variations (tomography) and the S-waves generated by P-wave interactions at sharp lithological boundaries in the lithosphere (receiver functions).

MT studies traditionally restrict themselves to linear transects on-land and use the natural electrical field variations as the structural probe. The development of ocean-bottom MT receivers is proceeding rapidly. The conductive ocean severely filters higher frequency natural sources and forces the use of active source methods to probe shallow ocean-bottom structure (4-6 km below the seafloor). Natural source ocean-bottom investigations are possible at longer period, but require correspondingly longer, more demanding ocean-bottom deployments.

The best-known study in which seismic and MT methods were combined is the investigation of the Cascade margin where the Juan de Fuca Plate subducts under Vancouver
Island (Figure 1). Here MT methods provided additional information into seismic structure by revealing a conductive layer which did not coincide with the inferred position of the subducted plate. The conductive layer lay above the inferred plate position in a zone of spatially discontinuous seismic reflectivity. The features suggest that fluids migrate upward from the plate into the wedge overlying the slab at 20-30 km depth. Above this same structure, a receiver function study using natural seismic sources located many of the same features seen in the active seismic source component, showing it provides information reconcilable with the other methods.

Seismic studies using natural earthquakes provide the most detailed picture of layering in subducted lithospheric slabs down to depths of 200 km. These employ S-to-P converted waves, P-to-S converted waves, and frequency-dependent apparent wave speeds to demonstrate the existence of low-velocity layers a few kilometers thick on or near the subducted plate surface. They provide long-distance average wave speed differences relative to the mantle and are not particularly sensitive to lateral variations on short scales of interest such as in regions of magma genesis. They do attest to thin layers in which seismic properties change over less than a kilometer. Thus layered structures persist at the slab surface to well below the expected depth of magma genesis.

Comparing wave speeds in the low velocity layers with those of metamorphic rock assemblages anticipated in subducted lithosphere yields some constraints on layer composition and slab interface temperature. They arise due to the strong differences in seismic wave speeds found in subduction zone metamorphic facies. Seismic wave speeds in an eclogite layer or one made of metastably-persisting gabbro conflicts with the observational data. They may be adequately explained by metastably-persisting blueschist, however, suggesting that slab surface temperatures are low. Tonga-Kermadec is an exception to the rule that wave speeds in slab layers are lower than the mantle. More than one mechanism must be responsible for velocity layering in slabs.

There is scope for broader integration of MT and seismological deployments. One suitable target is the nature of slab layering. Due to the low resistivity of fluid or weak zones, MT may be possible to use to identify whether such zones exist down to 200 km. The time scales for long-term MT deployments and natural source seismic investigations are similar (10^4-10^7 seconds), and it seems worthwhile to exploit the similarity of time scale and data acquisition rates to extend the knowledge of the subsurface.

References.


Figure 1. Figure from Jones (1998) after Hyndman (1988) showing seismic and resistivity structure along a transect across Vancouver Island. The conductive layer lies over the subducted Juan de Fuca plate.