MAGNETOTELLURIC SOUNDING OF TERRESTRIAL PLANET AND SATELLITE INTERIORS.
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Introduction. Electromagnetic (EM) sounding of the Moon and the Galilean satellites has yielded important insights on the interior structures of these bodies [e.g., 1-7]. These investigations measured magnetic fields only and required auxiliary information on the nature of the source fields. The magnetotelluric (MT) method, by measuring both magnetic and *electric* fields, can perform soundings from a single platform, without specific information on source fields. MT soundings can be performed from the ground, the air, or close orbit. Investigation depths vary depending on the spectra of source fields, but could span hundreds of meters to hundreds of kilometers. Greater investigation depths are enabled by the lower electrical conductivities of the outer shells of most targets due to the lack of liquid water. Component technology is mature and comparatively low-mass and low-cost. MT would be a useful addition to many missions investigating the solid bodies of the Solar System.

Terrestrial and Planetary EM Sounding. Time-varying natural or artificial EM fields induce eddy currents in planetary interiors, whose secondary EM fields are detected at or above the surface. These secondary fields shield the deeper interior according to the skin-depth effect, so that EM fields fall to 1/e amplitude over depth δ (km) = $0.5\sqrt{\rho/f}$, where ρ is the resistivity and f is the frequency. EM sounding exploits the skin-depth effect by using measurements over a range of frequency to reconstruct resistivity over a range of depth [8,9]. Natural EM signals (magnetospheric pulsations, ionospheric currents, lightning) are used instead of transmitters at the low frequencies necessary to penetrate kilometers to hundreds of kms into the Earth.

The fundamental quantity that must be derived in any sounding is the frequency-dependent EM impedance Z, and it is the variety of approaches to Z that lead to more individual techniques in EM than in any other geophysical method [e.g., 9]. The impedance is related to the apparent resistivity ρ_a—the most commonly used parameter beause of its dimensional analog to true resistivity—as $\rho_a = Z^2/\mu\omega$, where μ is the permittivity and ω is the angular frequency. Alternative EM response parameters such as the admittance or the transfer function can also be related to Z [10,11].

Two known quantities are necessary to determine the impedance, e.g., Ohm's Law $Z = V/I$. One of those quantities is nearly always the magnetic field near the target, i.e., the sum of source + induced magnetic fields. In a variety of *Transfer-Function* methods, the second known quantity is the source magnetic field. This is straightforward for an artificial transmitter. In a few special cases of accurately characterized natural signals—the Earth's ring current [12] or the time variation introduced by the motion of the Galilean satellites in Jupiter's main field [7]—the source can be specified a priori. A single magnetometer is also sufficient at very low frequencies where the source wavelength can be specified (e.g., a diurnal variation) or where the target can be approximated as a perfect conductor [e.g., 6]. Alternatively, the source magnetic field must be measured by a second, distant spacecraft, as was done for Apollo-era lunar soundings [1-5].

The *Geomagnetic Depth Sounding* (GDS) method uses surface arrays of magnetometers to determine impedance from the ratio of the vertical magnetic field to the magnitude of the horizontal magnetic-field gradient [13]. Because the wavelength in the ground $\lambda = 2\pi\delta$, GDS arrays require station spacings comparable to the skin depth in order to resolve the relevant horizontal wave structure. Therefore a sparse, global magnetometer network will only resolve planetary-scale structure, and dense regional arrays that could resolve shallow structure would be costly.

Magnetotelluric Method. A different approach offers a complete shallow-to-deep sounding from a single station, without specific knowledge of the source field. The *magnetotelluric method* (MT) uses orthogonal horizontal components of the local electric (E) and magnetic (B) fields to form $Z = \mu|E/B|$ [14,15]. The required second piece of information therefore is E. MT has vastly outpaced GDS in terrestrial exploration in recent decades because of its simplicity. Arrays are widely used, but only to provide more rapid geographic coverage and internal cross-checks among stations (remote reference). At planetary scales, the MT plane-wave response can be transformed to spherical geometry [10,11]. At small scale, MT naturally provides spatially independent measurements with horizontal resolution comparable to the EM skin depth. Therefore a single station can determine local-to-global structure, and multiple and/or mobile stations can assess lateral heterogeneity.

Magnetic-Field Measurement. MT requires vector measurements, so search-coils are appropriate for high frequencies and fluxgates for low frequencies. Because the horizontal fields measured by MT are not strongly affected by the subsurface, the source-field strength can be taken to determine the sensitivity requirement. This will typically be several pT near 1 Hz.

Electric-Field Measurement. Coupling of the voltage probes to the environment is effectively through a parallel RC circuit, so source impedances vary as 1/frequency in the high-frequency capacitive branch and are constant in the low-frequency galvanic branch. The crossover frequency decreases with resistivity. Therefore ground-contacting electrodes in the conductive Earth are nearly always in galvanic contact. Capacitive coupling is exploited in some high-frequency, small-scale electrical surveys [16] but not in low-frequency, large-scale MT. We have been working under PIDDP and MIDP [e.g., 17] to develop electrometers that can function capacitively at low frequency in resistive environments. Voltage-probe separations must also be restricted to several meters for planetary applications. Crossover frequencies are 1-10 Hz for most environments: even on Mars and in space there is sufficient ionization (atmosphere or plasma) to ensure galvanic coupling at low frequency. The extremely resistive atmosphere of Venus (like Earth) pushes the crossover down, increasing the impedance. The electric field increases with resistivity, $E = B\sqrt{\rho_a \omega/\mu_0}$, so larger electric fields will be present for most extraterrestrial applications than for the Earth. Sensitivity requirements are ~μV/m near 1 Hz.

Measurement Geometries. Although a ground station is perhaps the most natural measurement mode, MT can also be performed in principle in air or space. The rule of thumb is that the altitude cannot exceed about half the skin depth at the highest frequency used, or else the surface looks like a perfect conductor and there is no information about the subsurface. An additional constraint for a moving platform is that 10-100 periods should be recorded within a horizontal resolution element, again about a skin depth: this places a lower limit to useful frequency. Finally, MT cannot probe through a highly conducting ionosphere unless its properties are extraordinarily well characterized.

Data Processing and Interpretation. Complex impedances, formed in the spectral domain using least squares, are transformed to the real quantities apparent resistivity and phase. Standard nonlinear inversions recover resistivity as a function of depth. Because resistivity is a strong function of temperature, deep EM sounding is a window into thermal conditions of the interior and is a surrogate for heat flow. Interconnected, conductive graphite or iron-bearing minerals can influence bulk EM properties in special environments, but in general the resistivity of terrestrial planet deep interiors is dominated by mobile charge defects introduced by trace quantities of trivalent cations (Al, Fe) and absorbed H_2O [18-20]. Impurities in ice and the presence of salt hydrates will dominate the conductivity of icy moons [21].

Applications. *Moon.* Prior EM soundings of the Moon have placed upper limits on core size, determined the abundance of free iron in the upper mantle, and constrained the mantle temperature structure and global thermal evolution [e.g., 1-6]. The next generation of measurements can focus on higher frequeny waves present in the solar wind in order to probe the nature of the upper mantle and crust [22], specifically, the mantle composition [e.g., 23] and the magnitude and spatial heterogeneity in heat flow. MT will be insensitive to source structure that could affect the transfer function. Special care must be taken to characterize the plasma environment to remove any noninductive effects, however.

Mars. MT can determine the presence or absence of groundwater at depths of hundreds of meters to tens of kilometers [24]. The thickness of the cryosphere is an indicator of thermal gradient. Long-term measurements from a surface station can probe through any conductive middle and lower crust of Mars and assess the upper-mantle conductivity.

Venus. High-priority questions can be answered with MT [25]: What is the thickness of the lithosphere and how did it influence geodynamical evolution? What is the thickness of the crust and how did it differentiate? What is the water content of the mantle?

Mercury, Outer Planet Satellites. Landed or orbital platforms can determine the lithospheric/shell structures as well as the depths to internal oceans on the icy satellites. Balloons are attractive MT platforms for Venus and Titan.

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