MAGNETOTELLURIC SOUNDING OF THE INTERIOR OF VENUS. R. E. Grimm[1] and G. T. Delory[2], [1]Department of Space Studies, Southwest Research Institute, 1050 Walnut St. #300, Boulder, CO 80302 (grimm@boulder.swri.edu); [2]Space Sciences Laboratory, Univ. of California, Berkeley, gdelory@ssl.berkeley.edu.

Introduction. Electromagnetic (EM) sounding has been widely used to reveal earth structure from depths of meters to hundreds of kilometers [e.g. 1-5] and has also probed the deep interiors of the Moon [e.g. 6-9] and the Galilean satellites [10]. The magnetotelluric method [MT: 11,12] enables natural-source soundings from a single platform, without any external measurements or prior knowledge as was required for previous planetary EM applications. MT can in principle be performed anywhere within about 1 skin depth of the target, allowing measurements from the ground, the air, or in space (as long as the last is within any shielding ionosphere or current sheath). For Venus, MT measurements from the surface or an aerial platform can reveal lithospheric and crustal structure, providing key constraints on thermal and geodynamic evolution.

Terrestrial and Planetary EM Sounding. Time-varying natural or artificial EM source 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 [2,11,12]. The EM response is often expressed as the apparent resistivity ρ_a, the resistivity of a halfspace giving the same signal as the structure under test. Alternative EM response parameters such as the impedance, admittance or the transfer function can also be expressed as ρ_a [9,11-13].

To date, planetary EM sounding has used one or more magnetometers by (1) a priori knowledge of the amplitude and phase of the source field, such as the time variation introduced byt he motion of the Galilean satellites through Jupiter's main field [10], or (2) use of a second, distantly orbiting magnetometer to measure the source field, as was done using the Apollo 12 surface magnetometer and Explorer 35 spacecraft [6-9]. 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.

Magnetotelluric Method. A classical EM-exploration 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 (H) fields to form $\rho_a = (1/\mu\omega)(E/H)^2$, where μ is the permittivity and ω is the angular frequency [11,12]. At planetary scales, the MT plane-wave response can be transformed to spherical geometry [9,13]. At small scale, MT naurally provides spatially independent measurements with horizontal resolution comparable to the EM skin depth. Therefore a single station can determine global and local structure, and multiple and/or mobile stations can assess lateral heterogeneity.

Challenges associated with measurement and interpretation of the electric field include (1) Capacitive coupling of signals in regions of high resistivity; (2) Elimination of electrostatic effects from measurements made in plasmas; (3) Elimination of cross-contamination by high-amplitude vertical ("fair-weather") fields in planetary atmospheres, and (4) Erroneous "static shift" in ρ_a caused by spatial and temporal aliasing due to local conductors. Our PIDDP and MIDP programs have extended previous work in resistive environments [14] to develop high-impedance electrometers that enable capacitive measurements at low frequencies. Plasma effects can be assessed with a Langmuir probe and/or electrostatic analyzer. The time variation in vertical E-field is apparent in aerial surveys only and is largely due to measureable and modelable changes in platform attitude. Very large conductivity contrasts in the shallowest parts of the Earth that give rise to static shifts are not likely to be common in the resistive (cold and/or dry) outer portions of other solid bodies in the solar system. A mobile platform can average out static shift within the lateral footprint (again, about one skin depth).

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 [11,12]. 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 [4,5]. 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 electron hopping introduced by trace quantities of trivalent cations (Al, Fe) and by proton hopping introduced by absorbed H_2O [15-17].

Venus Science. EM sounding can address the following high-priority questions for Venus:

(1) *What is the thickness of the lithosphere?* Discounting a purely conductive-equilibrium endmember,

estimates of the asymptotic lithospheric thickness L_∞ vary from ~100 km for a mobile convective lid to ~500 km for a stagnant lid [18,19]. Determination of lithospheric thickness is pivotal to understanding how Venus loses its heat and hence understanding its geological history, particularly past and present tectonics and volcanism. Interior temperatures T and lithospheric thickness L can be constrained using laboratory data (Fig 1), as mentioned above. However, it is possible to determine L from just two measurement locations, without any supporting laboratory data, under the assumption that the topographic difference h_2-h_1 is purely due to lithospheric thermal isostasy. For a maximum EM penetration depth d, assume $T(d)$ = const and dT/dz = const in the mantle lithosphere: therefore $L_2/L_1 = d_2/d_1$. But from isostasy, $L_2 - L_1 = (h_2 - h_1)\rho_L/\Delta\rho$, where ρ_L is the mean lithospheric density and $\Delta\rho$ is the lithosphere-asthenosphere contrast. In practice, a least-squares solution involving all measurement locations would be performed, and the data should be restricted to lowlands where crustal-thickness variations are thought to be minimal [20]. Note that in neither approach does the EM sounding have to extend all the way through the lithosphere in order to determine its thickness.

(2) *What is the water content of the mantle?* The lower activation energy of hydrous olivine compared to dry olivine leads to a distinct slope in apparent resistivity vs. frequency (Fig. 1). Comparison with laboratory data [16] are necessary to quantify H_2O content, which should be resolvable at ppm levels and thence constrain degassing models [21].

(3) *What is the thickness of the crust*? The style of crustal accretion and destruction controls crustal thickness and composition [20]. Crustal thickness in non-highlands regions is thought to be relatively constant, but the mean value is still uncertain by up to a factor of 2 [20, 22]. Determination of crustal thickness in highlands can separate thermal/compositional isostasy [23].

Venus Environment and Application. Natural EM signals for Venus include solar-wind MHD oscillations <1 Hz [24] and lightning >1 Hz [25]. Electric fields are best measured at the surface, using long baselines from ballistic electrode deployment and exploiting partial galvanic coupling with the hot surface. Frequencies >10^{-2} Hz can be accurately characterized with a 1-hr lifetime. High-temperature magnetometer development is the critical path to surface measurements. Alternatively, in the benign environment at 50-60 km altitude, H-field measurements are simple but the E-field measurement is challenging due to the resistive atmosphere. For an aerial survey, the altitude must be less than ~1 skin depth, so depths >50 km could be targeted, but depths <50 km would lie in a blind zone. Aerial measurements would therefore be optimized for the mantle and able to detect only the thickest highlands crust. Long-lived, mobile measurements also offer the greatest data diversity, improving SNR and improving separation of the contributions of temperature and composition to subsurface resistivity.

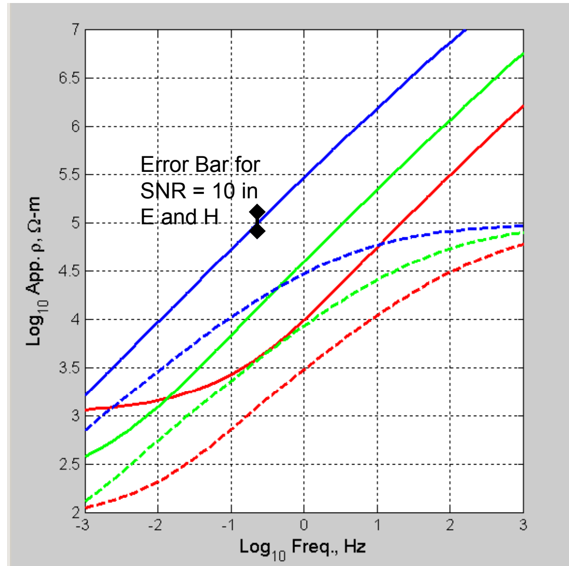

Figure 1. Forward modeling of MT signatures of different thermal structures and mantle compositions of Venus. Red: Incipient melting, L = 30 km, ΔT = 750 K. Blue: Mobile lid, L = 100 km, ΔT = 1000 K. Green: Stagnant lid, L = 500 km, ΔT = 1300 K. Solid: Dry olivine. Dash: Olivine with 1 ppm H_2O. Note separation of L curves >> error bar and different frequency slope of water-bearing mantle.

References. [1] McNeill, J.D. (1990) in *Geotech. Environ. Geophys*. (ed. S.H. Ward) Soc. Explor. Geophys. (SEG), Tulsa, p. 191. [2] Spies, B.R., and F.C. Frischknecht (1991) in *EM Meth. Appl. Geophys*., V. 2 (ed. M.S. Nabighian) SEG., Tulsa, p. 285. [3] Banks, R.J. (1969) *GJRAS, 17,* 457. [4] Boerner, D.E. et al. (1999) *Science, 283,* 668. [5] Jones, A.G. et al. (2005) *CJES, 42*, 1257. [6] Blank, J.L. and W.R. Sill (1969) *JGR, 74*, 736. [7] Dyal P. et al. (1977) *RGSP, 12*, 568. [8] Hood L.L. et al. (1982) *JGR, 87*, 5311. [9] Hobbs, B. et al. (1983), *JGR, 88*, B97-B102. [10] Khurana, K.K., et al. (1998), Nature, 395, 777. [11] Vozoff, K. (1991) in *EM Meth. Appl. Geophys*., Vol. 2 (ed. M.S. Nabighian) SEG, Tulsa, p. 641. [12] Simpson, F., and K. Barr *(2005) Practical Magnetotellurics*, Cambridge, 254 pp. [13] Weidelt, P. (1972), *Zeit. Geophys*., *38,* 257. [14] Wannamaker, P. et al. (1996) *GRL, 23*, 2983. [15] Huebner, J.S. et al. (1979) *JGR, 84*, 4652. [16] Wang D. et al. (2006) *Nature, 443*, 977. [17] Tyburczy, J.T. (2007) in *Treat. Geophys*., p. 631. [18] Kaula, W.M. and R.J. Phillips (1981) *GRL, 8,* 1187. [19] Reese, C.C. et al. (1998) *JGR, 103,* 13643. [20] Grimm R.E., and P.C. Hess (1997) in *Venus II* (eds. S. Bougher et al.), Univ. Ariz., p. 1205. [21] Namiki, S., and S. Solomon (1995) LPSC XXVI, 1029. [22] Konopliv, A.S. et al. (1999) *Icarus, 139,* 3. [23] Grimm, R.E. (1994) *Icarus, 112*, 89. [24] Luhmann, J.G. (1991) *JGR, 96*, 18831. [25] Russell, C.T. et al. (2007) *Nature, 450*, 661.