InSight: Single station broadband seismology for probing Mars' interior.

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Introduction: InSight is a proposed Discovery mission which will deliver a lander containing geophysical instrumentation, including a heat flow probe and a seismometer package, to Mars. The aim of this mission is to perform, for the first time, an in-situ investigation of the interior of a truly Earth- like planet other than our own, with the goal of understanding the formation and evolution of terrestrial planets through investigation of the interior structure and processes of Mars.

SEIS Instrumentation: One of the critical instrumentation components for the InSight mission will be the Seismic Experiment for Interior Structure (SEIS) [1]. SEIS comprises two sensor assemblies mounted on a leveling mechanism which will be deployed on the surface: a 3- axis very broad band (VBB) oblique seismometer within an evacuated sphere, and an independent 3-axis short period (SP) seismometer outside. The combination of the two sensor assemblies allows for high precision measurements over a very broad frequency band ($<10^{-9}$m/s^2/Hz$^{1/2}$ between 0.001 and 1 Hz for VBB and $<5\times10^{-9}$m/s^2/Hz$^{1/2}$ between 0.01 and 50 Hz for SP).

Such instrumentation will allow for an unprecedented view of the interior of Mars. However, since the proposed mission will have only a single lander and no network, we will not be able to apply traditional source location methods and will need to take advantage of single station approaches.

Expected sources of energy: A first-order scientific return of such a mission will be information on the seismic activity of the planet. We can, however, use estimates of Martian seismicity based on thermal calculations [2] or extrapolation of historical faulting [3, 4] to estimate data availability. Based on these estimates and the capabilities of the SEIS instrumentation, we anticipate being able to record body wave information for between 30 and 40 quakes of seismic moment greater than or equal to 10^{13}Nm during 2 years. Of these, we anticipate on the order of 4 to 5 events with seismic moment greater than or equal to 10^{16}Nm ($\sim M_W 4.7$), which will be large enough to allow for recording of at least 3 orbits of Rayleigh waves.

Multiple orbit surface wave data: When a single seismic station is available, the record of the successive surface wave trains generated by an event can be used to assess the spherically averaged phase and group velocities as a function of frequency, the epicentral distance, and the origin time of the event. On the Earth, the determination of seismic velocities along the great circle paths is a standard procedure, initiated in the 1950's and 1960's [5, 6], that has been widely used for several decades [e.g. 7, 8], and can be applied without any knowledge of the source location and origin time or correction for instrumental response. Using the R1, R2, and R3 notation for the successive minor and major arc wavetrains, and the signal-to-noise ratio expected for SEIS, R1 wavetrains generated by a Mw 4.7 event can be observed on the Earth at 5000 km epicentral distance in the 35-50 s period range, but R2 and R3 wavetrains are below the noise level (fig. 1). This is mainly due to the large associated travel distances (35 000 km and 45 000 km), respectively, and to the damping effect of the oceans. On Mars, the smaller planetary radius (about half of the Earth's) and the absence of oceans, result in a much weaker attenuation of the dispersed wavetrains in this period range. This should enable the SEIS seismometer to observe R2 and R3 wavetrains at all distances for marsquakes with moment magnitude greater than 4.7 (fig. 1). Considering a very conservative 20 s error on arrival time determinations, the precisions expected on great-circle surface-wave velocities and epicentral distances are expected within the ranges 1% and 200 km, respectively.

For these events where we are able to estimate epicentral distance from surface wave arrival times, we can make arrival time picks for body wave phases as well, allowing for the estimation of P and S travel time curves. If enough picks are available to fit a smooth curve to the travel time picks, we can use classical seismology techniques such as Herglotz-Wiechert inversion [e.g. 9, ch. 9] to create a smoothed velocity profile for the mantle. Even if the picks are too sparse for such an analytic approach, a few differential S-P times can be used in connection with the constraints from the Rayleigh wave dispersion characteristics to constrain simple mantle velocity models. With the mantle velocity constraints, any reflected phases observed in these events, such as PcP or ScS can then be used to constrain the core radius.

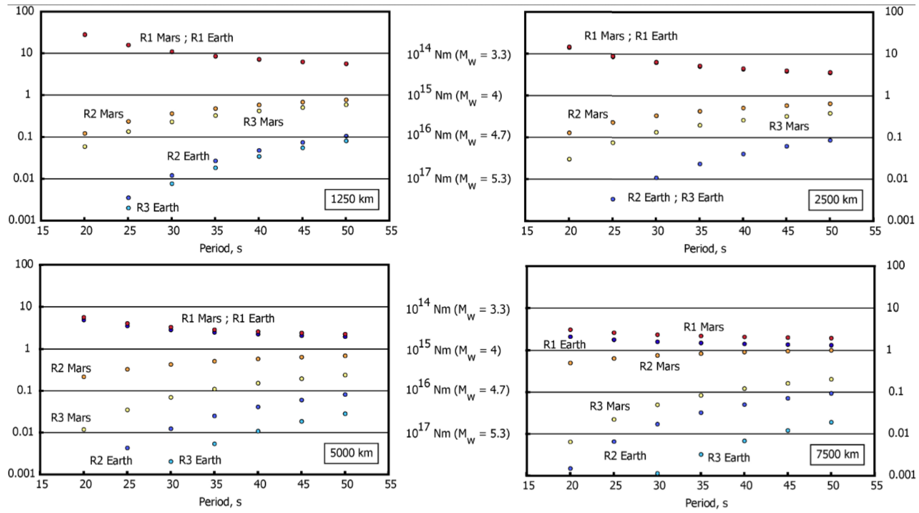

Figure 1: Amplitude of Rayleigh wave trains normalized by R1 amplitude at an epicentral distance of 90° on Earth (10,000 km). S/N numbers on y-axis are for $M_W 4$, but seismic moment labels in center of figure indicate the minimum amplitude that is required to observe a particular wave train with a S/N ratio of 1.

Other single-station approaches: For events where we may be able to observe some body wave or surface wave energy but not multiple orbits to estimate epicentral distance, we can still apply well-developed single-station approaches to constrain structure of the crust and upper mantle. In particular, the method of receiver functions, which only requires three component recordings of incoming P or S waves and no source information has been in use for decades [10] and has been applied to lunar data [11], should be able to constrain crustal thickness as well as possibly identifying upper mantle discontinuities. Further constraints on upper mantle structure may be possible from other approaches such as the analysis of the ellipticity of surface waveforms [12].

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