InSight: Using Earth data to demonstrate inversion techniques for 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. The chief science goal is understanding the formation and evolution of terrestrial planets through investigation of the interior structure and processes of Mars.

SEIS Instrumentation: The critical instrument for the InSight mission will be the Seismic Experiment for Interior Structure (SEIS) [1]. SEIS comprises two sensor assemblies 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 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, and these approaches should be tested with Earth data.

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 visible 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 ∼200 quakes of seismic moment greater than or equal to 10^{14}Nm during 2 years. Of these, we anticipate on the order of 8 events with seismic moment greater than or equal to 10^{16}Nm ($\sim M_\mathrm{W}$4.7), which will be large enough to allow for recording of at least 3 orbits of Rayleigh waves.

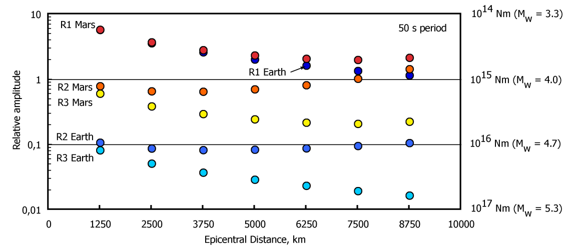

Figure 1: Amplitude of Rayleigh wave trains normalized by R1 amplitude at an epicentral distance of 90° on Earth (10,000 km). S/N numbers on y-axis are for M_W4, but seismic moment labels on right of figure indicate the minimum amplitude that is required to observe a particular wave train with a S/N ratio of 1.

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 [5] that has been widely used for several decades [e.g. 6, 7], and can be applied without any knowledge of the source location and origin time. 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 M_W 4.7 event can be observed on the Earth out to 10,000 km epicentral distance at 50 s period, but R2 and R3 wavetrains are below the noise level (fig. 1). This is mainly due to the large associated travel distances (30,000 km and 50,000 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, allowing observation of R2 and R3 wavetrains at all distances for marsquakes with moment magnitude greater than 4.7 (10^{16} Nm).

For these events where we are able to estimate epicentral distance from surface wave arrival times, we can

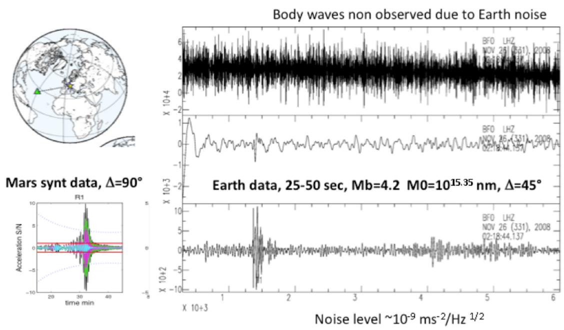

Figure 2: Example comparison of synthetic Martian R1 seismogram (lower left) with seismogram recorded at station BFO at a similar distance and seismic moment. R1 is clear in the 25-50 second pass band (lower right), but the body wave arrivals are obscured by oceanic microseismic noise in the unfiltered data (top right), which will not be present on Mars. The middle right panel shows longer period data filtered between 40 and 250 s, demonstrating that R1 is most clear in the 25-50 s band.

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. 8, 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.

Verification with Earth data: In order to test the feasibility of such a single-station approach, we present estimates of Earth structure using events recorded at a single low-noise station. One such station is Black Forest Observatory in Germany (fig. 2), which shows a similar R1 wave train to Martian synthetics. Because the larger radius of the Earth leads to smaller R2 and R3 amplitudes, we use larger events than expected on Mars to apply the techniques discussed here. Additionally, oceanic microseismic noise, which will not be present on Mars due to the lack of oceans, obscures body wave phases at lower magnitudes on Earth. We will present estimates of structure using varying quantities of data to demonstrate how well we can constrain structure with different rates of marsquakes.

Other single-station approaches: For events where we may be able to observe body or surface wave energy but not multiple orbits to estimate epicentral distance, we can 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 [e.g. 9] and has been applied to lunar data [10], can constrain crust and shallow mantle structure. Large events may also allow for observation of normal modes, which can then constrain structure throughout the planet.

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