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A Pre-Landing Assessment of Regolith Properties at the InSight Landing Site

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38 Abstract

39 This article discusses relevant physical properties of the regolith at the Mars InSight landing site as
40 understood prior to landing of the spacecraft. InSight will land in the northern lowland plains of Mars,
41 close to the equator, where the regolith is estimated to be ≥ 3 -5 m thick. These investigations of
42 physical properties have relied on data collected from Mars orbital measurements, previously collected
43 lander and rover data, results of studies of data and samples from Apollo lunar missions, laboratory
44 measurements on regolith simulants, and theoretical studies. The investigations include changes in
45 properties with depth and temperature. Mechanical properties investigated include density, grain-size
46 distribution, cohesion, and angle of internal friction. Thermophysical properties include thermal inertia,
47 surface emissivity and albedo, thermal conductivity and diffusivity, and specific heat. Regolith elastic
48 properties not only include parameters that control seismic wave velocities in the immediate vicinity of
49 the InSight lander but also coupling of the lander and other potential noise sources to the InSight
50 broadband seismometer. The related properties include Poisson's ratio, P- and S-wave velocities,
51 Young's modulus, and seismic attenuation. Finally, mass diffusivity was investigated to estimate gas
52 movements in the regolith driven by atmospheric pressure changes. Physical properties presented here
53 are all to some degree speculative. However, they form a basis for interpretation of the early data to be
54 returned from the InSight mission.

55

56 Keywords

57 Mars, Regolith, Physical Properties, InSight Landing Site

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95 **Introduction**

96 The InSight mission is the first dedicated geophysical mission to another planet. InSight (Interior
97 Exploration using Seismic Investigations, Geodesy and Heat Transport) will place a single geophysical
98 lander on Mars to study its deep interior and to provide information relevant to the fundamental
99 processes of terrestrial planet formation and evolution (Banerdt et al., 2013). This article discusses the
100 physical properties of the Mars regolith at the InSight landing site based upon information available
101 approximately one year prior to launch, and eighteen months prior to touchdown of the InSight lander.
102 The InSight mission represents many years of engineering and scientific design and preparation, based
103 to some degree on the properties of the regolith at the landing site. Most of the scientific data to be
104 collected by instruments on the InSight lander will be filtered by the regolith in the immediate vicinity of
105 the landing site. Therefore to design these instruments and to make realistic predictions of the range of
106 data characteristics that should be recorded by the instruments, a model of the physical properties of
107 the landing site regolith has been required. As the science team approaches the final stages of
108 preparation for first data return from the InSight Mission, we saw benefit in using a consistent set of
109 regolith physical property values for any required data processing and early publications across the
110 project. At least some of these property values will be revised at a later date with new data from the
111 InSight instruments.

112 The InSight lander is based on a lander used for the successful Phoenix mission that was launched to
113 Mars on August 4, 2007 and investigated near-surface ice in the Martian Arctic (Smith et al., 2009).
114 Scientific instruments on the Phoenix lander have been replaced by a SEIS seismometer (Seismic
115 Experiment for Interior Structure) comprising a very broad-band (VBB) and a short period (SP)
116 seismometer that will be placed on the surface of Mars, a heat-flow probe with an internal hammer
117 mechanism that will hammer itself into the Martian regolith with an accompanying radiometer to
118 determine the radiative surface temperature of the regolith close to the lander, and a precision tracking
119 system. Additional instruments on the lander will measure orbital and local atmospheric parameters of
120 Mars. Some regolith properties, such as radioactivity and magnetic properties have been omitted in this
121 discussion because they were either not pertinent to the InSight Mission instruments or they lacked
122 data at the regolith scale.

123 The InSight landing site is shown on a portion of Mars topography in Figure 1. The general landing
124 area was chosen for basic operational reasons of being close to the equator for year-round solar power

125 for the lander and smooth topography for the landing site. More specific details of landing site selection
126 are given in the [Landing Site Overview](#) in section 2.1 below. Mars has two basic terrains, smooth
127 northern lowland plains (“planitia”) and southern cratered highlands (“terra”), separated by the
128 dichotomy boundary. Four geologic eras have been assigned to terrains on Mars based on crater
129 densities: Pre-Noachian, 4.5 – 4.1 Ga; Noachian, 4.1 – 3.7 Ga; Hesperian, 3.7 – 3.0 Ga; and Amazonian,
130 3.0 Ga – present. The landing site is in lowlands terrain of Early Hesperian or younger age, just north of
131 the dichotomy boundary.

132 == Figure 1 about here ==

133 Following this introduction is a description of the regolith at the landing site including the criteria and
134 process of landing site selection. This section is followed by four regolith physical property sections:
135 Regolith Soil Mechanical Properties; Regolith Thermo-Physical Properties; Regolith Elastic Properties;
136 and Mass Diffusivity. The paper closes with a summary and conclusions section. Sections were
137 contributed by different authors or groups of authors according to their specialty. We have endeavored
138 to make the document flow as smoothly as possible, but it is primarily an informational article.
139 However, what the paper lacks in style we hope that it contributes in utility.

140 **1. Regolith at the InSight Landing Site**

141 This section describes properties of the regolith essential for safe landing and operation of the
142 spacecraft and instrument deployment.

143 **1.1. Landing Site Overview**

144 InSight will land in western Elysium Planitia on Hesperian plains just north of the dichotomy
145 boundary (Golombek et al., 2017). This location satisfies the three dominant landing site engineering
146 constraints, which are latitude (3°N-5°N), elevation (<-2.5 km with respect to the MOLA geoid), and a
147 large smooth, flat surface to place a 130 km by 27 km landing ellipse. Other engineering constraints that
148 are relevant to the geologic setting include: 1) a load bearing, radar reflective surface with thermal
149 inertia >100–140 J/(m² K s^{1/2}), slopes <15° and rock abundance <10% for safe landing and instrument
150 deployment, and a broken up regolith >3 m thick to facilitate deployment of the heat flow probe
151 (Golombek et al., 2017).

152 The InSight landing ellipse is located on smooth plains with Noachian highlands to the south and
153 west, a ridge of Medusae Fossae Formation to the southeast and very young lavas from Athabasca
154 Valles to the east (Golombek et al., 2017). The ellipse is located at 4.5°N, 135.9°E, about 540 km north
155 of the Mars Science Laboratory landing site. The plains surface on which the InSight ellipse is located is
156 mapped as Early Hesperian transition unit (eHt) by Tanaka et al. (2014) in the global geologic map of
157 Mars, which could be sedimentary or volcanic. A volcanic interpretation of the plains is supported by: 1)
158 the presence of rocks in the ejecta of fresh craters ~0.4 to 20 km diameter suggesting a strong
159 competent layer ~4 to 200 m deep with weaker material above and below (e.g., Golombek et al., 2013;
160 Catling et al., 2011, 2012; Warner et al., 2017); 2) exposures of strong, jointed bedrock overlain by ~10
161 m of relatively fine grained regolith in nearby Hephaestus Fossae in southern Utopia Planitia at 21.9°N,
162 122.0°E (Golombek et al., 2013, 2017); 3) platy and smooth Late Hesperian to Early Amazonian lava
163 flows up to 200 m thick mapped in 6 m/pixel visible images south of the landing site (Ansan et al., 2015);
164 and 4) the presence of wrinkle ridges, which have been interpreted to be fault-propagation folds, in
165 which slip on thrust faults at depth is accommodated by asymmetric folding in strong, but weakly
166 bonded layered material (such as basalt flows) near the surface (e.g., Mueller and Golombek, 2004;
167 Golombek and Phillips, 2010).

168 The landing ellipse has very low rock abundance (Golombek et al., 2017). Most rocks at the landing
169 site are concentrated around rocky ejecta craters larger than 30 to 200 m diameter, but not around
170 similarly fresh smaller craters (Golombek et al., 2013, 2017). Because ejecta is sourced from shallow
171 depths, ~0.08 times the diameter of the crater (Melosh, 1989), and based on the assumption that the
172 surface morphology is fresh and not highly eroded, the onset diameter of rocky ejecta craters has been
173 used to map the thickness of the broken up regolith. Results indicate a regolith that is 3-17 m thick
174 (Warner et al., 2014, 2016, 2017), that grades into large blocky ejecta over strong intact basalts
175 (Golombek et al., 2013, 2017). Because fresh craters larger than 2 km do not have rocky ejecta, material
176 below the basalts at ~200 m depth is likely weakly bonded sediments.

177 Surficial thermophysical properties of the landing site indicate that the soil that makes up the surface
178 materials is similar to common weakly bonded soils on Earth and conducive to penetration by the heat
179 flow probe (Golombek et al., 2017). The thermal inertia of the landing ellipse is about $200 \text{ J}/(\text{m}^2 \text{ K s}^{1/2})$,
180 the albedo is 0.25, and dust cover index is 0.94 (see Section 4.2, and Golombek et al., 2017).
181 Comparison with the thermal inertias of previous landing sites and the soils at these sites (Golombek et
182 al., 2008a) suggests the InSight landing site surfaces are composed of cohesionless sand or low cohesion

183 soils (cohesions of less than a few kPa, angle of internal friction of 30-40°), with bulk densities of ~1000
184 to 1600 kg/m³, particle sizes of ~150-250 μm (fine sand), that extend to a depth of at least several tens
185 of centimeters, and with surficial dust layer less than 1–2 mm thick (Golombek et al., 2017).

186 The albedo and dust cover index are similar to dusty and low-rock abundance portions of the Gusev
187 cratered plains, which are Hesperian lava flows with an impact generated regolith, modified by eolian
188 processes (Golombek et al., 2006). Mapping of surface terrains in high-resolution images of the InSight
189 landing site and surrounding areas, shows these terrains are similarly Hesperian lava flows with an
190 impact generated regolith modified by eolian processes (Golombek et al., 2017; Warner et al., 2017).

191 An exposed escarpment of nearby Hephaestus Fossae (Figure 2) shows this near surface structure
192 with ~10 m thick, relatively fine grained regolith, that grades into coarse, blocky ejecta with meter to
193 ten-meter scale boulders that overlies strong, jointed bedrock. The grading of finer grained regolith into
194 coarser, blocky ejecta is exactly what would be expected for a surface impacted by craters with a steeply
195 dipping negative power-law size distribution in which smaller impacts vastly outnumber larger impacts
196 that would excavate more deeply beneath the surface (e.g., Shoemaker and Morris, 1969; Hartmann et
197 al., 2001; Wilcox et al., 2005).

198 == Figure 2 about here ==

199

200 **1.2. Rock Abundance**

201 The contrast between measurements of thermal emission from the surface at various wavelengths
202 using the Viking Orbiter Infrared Thermal Mapper (IRTM) and the Mars Global Surveyor spacecraft
203 Thermal Emission Spectrometer (TES) data have been used to determine the rock abundance (the
204 fractional area covered by high thermal inertia rocky material) at about 60 and 8 km/pixel scales
205 (Christensen, 1986; Nowicki and Christensen, 2007). With the rock abundance and the bulk thermal
206 inertia, the thermal inertia of the remaining soil, referred to as the fine-component thermal inertia
207 (Kieffer et al., 1977), has also been determined (Christensen, 1986; Nowicki and Christensen, 2007).
208 Rock abundance estimated from thermal differencing is 4% and 9% for IRTM pixels of ~60 km
209 (Christensen, 1986) and around 4% (1%–7%) for TES pixels of ~8 km (Nowicki and Christensen, 2007) in
210 the landing ellipse. Because the thermal differencing estimates of rock abundance are relatively low for

211 this area (Christensen, 1986; Nowicki and Christensen, 2007), the fine component thermal inertia is only
212 slightly lower than the bulk thermal inertia.

213 Rock abundance measured from shadows in HiRISE images fit to model exponential cumulative
214 fractional area versus diameter curves in 150 m bins (Golombek et al., 2008b, 2012) also indicate a very
215 low average rock abundance of 1-2% for the InSight landing site (Golombek et al., 2017), although rock
216 abundance can increase to ~35% around rocky ejecta craters. Fragmentation theory in which the
217 particle size distribution is described by a negative binomial function (Charalambous, 2014) was applied
218 to the InSight landing site using cratering size-frequency measurements to derive a synthesized regolith
219 with a size-frequency distribution similar to the exponential model for ~2-6% rock abundance
220 (Charalambous et al., 2011; Golombek et al., 2017). The measurements and models of rock abundance
221 combined with the thermal inertia observations all indicate a relatively fine-grained regolith with low
222 rock abundance in the upper 5 m of the regolith at the landing site.

223 **1.3.Regolith Structure Summary**

224 In summary, the upper 5 m of regolith at the landing site are expected to be dominantly composed of
225 nearly cohesionless fine basaltic sand, which contains few rocks. The regolith was produced by impact
226 gardening of basalt flows with eolian sorting and transport of the sand. In contrast with lunar regolith,
227 the sand grains are rounded to sub-rounded by saltation (e.g., McGlynn et al., 2011). With increasing
228 depth, larger particles and rocks are expected to become more plentiful until the upper, relatively fine-
229 grained regolith grades into a coarse-grained breccia or blocky ejecta that overlies fractured basalt
230 flows. In addition, with increasing depth the effects of impact decreases and basalt would likely be less
231 fractured. Below ~200 m basalt would transition to sediments or weakly bonded sedimentary rocks.

232

233 **2. Regolith Soil Mechanical Properties**

234

235 **2.1 Introduction**

236 The parameters used to characterize the mechanical properties of the regolith at the InSight landing
237 site are considered in this section. They are also summarized in a table in the Appendix.

238 The Martian regolith is expected to be a complex mix of weathered, indurated, and windblown
239 material (e.g., Putzig and Mellon, 2007), and apart from engineering safety considerations, the InSight
240 landing site was chosen to facilitate penetration of the HP³ thermal probe to a depth of 3-5 m into a
241 column of fragmented regolith (Golombek et al., 2017). Comparison with data from other landed
242 missions and orbiters indicates that the regolith is largely cohesionless, has angle of internal friction
243 close to that of sand (30-40°), and particles are expected to be rounded due to erosion by wind. Indeed,
244 eolian activity on Mars has occurred throughout geologic time. The surface layer has been subjected to
245 eolian activity and impacts: after each impact sand size grains have been saltated and rounded and
246 sorted and the entire column of material has rounded (sub-rounded) grains. As such, the region may be
247 viewed as an eolian deposit which may be the result of potentially several inflation and deflation
248 periods. Given the values of thermal inertia (200 J/(m² K¹ s^{1/2})), albedo (0.25) and dust cover index
249 (0.94) in the InSight landing place, and based on comparison with the thermal inertias of previous
250 landing sites, the InSight surfaces are composed of cohesionless sand or low cohesion soils with particle
251 sizes of ~0.15-0.25 mm (fine sand) (Golombek et al., 2008a, 2017).

252

253 **2.2 Density**

254 Physical properties of regoliths, such as thermal conductivity, seismic velocity, penetration
255 resistance, shear strength, compressibility and dielectric constant, depend on bulk density, which
256 depends on grain size distribution, grain shape, particle surface texture and grain arrangement (Carrier
257 et al., 1973). In dust powders, repulsive effects of electrostatic forces can result in densities as low as
258 1000 kg/m³; in fine sand, inter-particle forces are mainly governed by gravity and inter-granular friction,
259 resulting in higher densities. However, it is likely that the lower gravity on Mars could result in looser
260 arrangements of grains of same shape and size distribution, compared to the gravity on the earth.
261 Possible values of the regolith density can be further estimated by considering typical features of
262 granular assemblies and sands, together with the physical properties of some terrestrial sands and
263 regolith simulants (Mojave simulant, Eifelsand, and Mars Soil Simulant-D; Delage et al., 2017). A simple
264 illustration providing first order estimates can be obtained from geometrical considerations of
265 arrangements of spherical particles of the same diameter. In the densest possible arrangement
266 (tetrahedral), with a minimum void ratio $e_{min} = 0.351$, with terrestrial sands, often composed of quartz
267 grains with a density of 2670 kg/m³, this value corresponds to a maximum bulk density of 1980 kg/m³, a

268 high density for (non-basaltic) terrestrial sands. For basaltic sands, as on Mars and in some areas on the
269 earth, the corresponding density would be 2230 kg/m^3 with a grain density of 3310 kg/m^3 for basalt.
270 Conversely, the loosest possible assembly of spheres (simple cubic) has a maximum void ratio $e_{max} =$
271 0.908 , yielding a minimum bulk density of 1400 kg/m^3 for quartz sands and of 1580 kg/m^3 for basaltic
272 sands. For non-spherical grain shapes, other configurations are possible. For example, elongated grains,
273 with aspect ratios significantly different from one, may exhibit rotational interlocking, particles resting
274 against each other building bridges that increase void space. Limited overburden pressure can prevent
275 particles from rotating and form statically stable regimes, supported in the low gravity of Mars, and
276 especially prevalent in particle packages that have not be subject to strong external loading. Once
277 loaded or subject to vibration, these packages will tend to increase in density.

278 On the Moon, regolith density drastically increases at depths below 20 cm. This increase has been
279 attributed to the effects of continuing small meteoroid impacts, not filtered by an atmosphere as on
280 Mars. Small impacts generate a loose, stirred-up surface while at the same time densifying the
281 underlying soil (Carrier et al., 1973). Details of this process are not fully understood (Heiken et al.,
282 1991), but best estimates for typical average densities are 1450 to 1550 kg m^3 at depths between 0 and
283 15 cm and 1690 to 1790 kg/m^3 at depths between 30 and 60 cm . In addition, analyses of the heat flow
284 experiment data emplaced at the Apollo 15 and 17 sites indicates that the bulk density must be
285 approximately 1300 kg/m^3 at the surface and must rise steeply in the upper few centimeters in order to
286 be consistent with nighttime surface temperature data (Keihm et al., 1973; Keihm and Langseth, 1973,
287 1975; Langseth et al., 1976). The situation is, however, quite different on Mars because
288 micrometeorites are stopped by the atmosphere. The primary shallow processes are wind transport
289 and saltation of sand-size particles.

290 In natural sands, a non-uniform grain size distribution provides denser arrangements, with smaller
291 grains filling voids between larger grains. Irregular angular grains allow for looser packing than
292 spherical grains. This is expected to be the case for the InSight landing site, with surface densities
293 estimated to be around 1300 kg m^{-3} (see below). Bolton (1986) provided the minimum (e_{min}) and
294 maximum (e_{max}) void ratios and densities of a series of terrestrial sands. The loosest sands were two
295 river sands (Welland River, Canada, and Chattahoochee River, USA) with bulk densities of 1390 and 1290
296 kg/m^3 , respectively. Note that river sands are known to be rounded due to transportation in water.
297 Sand on Mars is rounded during saltation (McGlynn et al., 2011). Both the minimum (1290 kg/m^3) and
298 maximum (1910 kg/m^3) densities provided by Bolton (1986) are not too far from densities obtained from

299 simple geometrical considerations on the ideal granular arrangements of spheres. In addition,
300 observations made by previous landers and rovers also showed bulk densities in the range of 1100–1300
301 kg/m³ and 1150 ± 150 kg/m³ for surficial sand and sandy soil deposits (see, e.g., Golombek et al.,
302 2008a; Herkenhoff et al., 2008, and references therein). Based on the fact that surface thermal inertia
303 values are most compatible with a sand to crusty-cloddy soil deposits (Golombek et al., 2008a) and given
304 the above considerations on terrestrial sands, the current best estimate for the regolith surface density
305 is close to 1300 kg/m³. In addition, a friction angle of about 30° would also correspond to this density
306 range (Delage et al., 2017).

307 In general, density is expected to increase with depth as a function of overburden pressure following
308 an exponential relation (e.g., Robinson and Gluyas, 1992; Revil et al., 2002), but compressibility of Mars
309 simulant material was found to be small, with an increase in density of around 20 kg/m³ from the
310 surface to 5 m depth (Delage et al., 2017), such that this effect can generally be neglected for the depth
311 range relevant here. Regolith particles on Mars initially originate from the comminution caused by
312 impacts on the surface, prior to being affected by eolian transportation and saltation that result in
313 reducing their initial angularity to produce rounded or sub-rounded sorted grains. While repeated
314 excavation, breakup, and movement by wind would result in a rather loose packing of grains,
315 subsequent vibrational compaction due to, e.g., seismic events may compact the soil to significant
316 depth, as is observed on the Moon (Carrier et al., 1973, 1974; Heiken et al., 1991). In addition, saltation
317 of grains during the soil deposition can be a high energy process and compact the soil, and relative
318 densities in excess of 90% have been observed in accretional deposits on terrestrial sand dunes
319 (Denekamp and Tsur-Lavie, 1981). Therefore, a model of regolith density for the InSight landing site
320 should allow for some compaction to be present.

321 Regolith structure may locally deviate from the model proposed above in regions where craters have
322 been filled with fine grained material due to eolian activity. This has been observed, for example, in the
323 Gusev plains, where craters with diameters between 20 and 100 m are abundant in all stages of erosion
324 (Golombek et al., 2006). Given a depth to diameter ratio of typically 0.2 for simple craters, filling by fine
325 grained material could provide lens of dominantly sand-sized material in the subsurface that have not
326 been mixed with rocks or other material by subsequent impacts.

327 To describe the lunar density data, a hyperbolic density relationship was established which
328 reasonably reproduces densities to a depth of 3 m. However, this description is not based on any

329 physical model. Rather, it was chosen because linearly, superlinearly, or exponentially increasing
330 profiles yield unrealistic values at the surface or at larger depths (Heiken et al., 1991), although they also
331 fit the available data. In its general form, density may then be written as:

$$\rho(z) = \rho_{inf} \frac{A + z}{B + z} \quad (1)$$

332 where $\rho(z)$ is density ρ as a function of depth, ρ_{inf} is the density at depth and z is the depth below the
333 Martian surface in meters. A and B are constants with the dimensions of length that describe the
334 chosen density profile, and example coefficients corresponding to the cases shown in Figure 3 are given
335 in Table 1. As a reference, a surface density of 1300 kg/m³ seems to be most compatible with the
336 available constraints, and three different compaction models are shown. If void ratios between $e_{min} =$
337 0.75 and $e_{max} = 1.5$ are assumed in accordance with measurements on Mars regolith analogue material
338 (Vrettos et al., 2014), relative densities between 0.6 (moderately compacted) and >0.9 (densely
339 compacted) are obtained at 5 m depth.

340 == Table 1 about here ==

341 == Figure 3 about here ==

342

343 **2.3 Cohesion**

344 Cohesion, a component of the shear strength, of surface materials on Mars has been determined
345 from soil mechanics experiments performed by arms and scoops on fixed landers and by the interaction
346 of wheels of rovers with surface materials by rovers. The two Viking landers and the Phoenix lander had
347 arms that trenched surface materials while monitoring motor currents to yield force, and imaging
348 systems to observe the deformed materials (Moore et al., 1977, 1987; Shaw et al., 2009). The Mars
349 Pathfinder rover, Sojourner, the two Mars Exploration Rovers, Spirit and Opportunity, and the Mars
350 Science Laboratory rover, Curiosity, performed wheel trenching and terramechanics experiments, while
351 monitoring motor currents to derive wheel torques, and imaged the deformed materials (Moore et al.,
352 1999; Herkenhoff et al., 2008; Sullivan et al., 2011; Arvidson et al., 2014). These experiments
353 determined basic soil mechanics measurements of cohesion and angle of internal friction. Imaging and
354 thermophysical properties and other relations were used to measure or constrain the particle size of the

355 soils and the bulk density (e.g., Moore and Jakosky, 1989; Christensen and Moore, 1992; Herkenhoff et
356 al., 2008; Golombek et al., 2008a).

357 Results of these experiments revealed four probable different soil deposits on Mars based on their
358 mechanical properties and likely means of formation (e.g., Golombek et al., 2008a). Two types of
359 deposits that appear to have been deposited by the wind were found at the landing sites. 1) Bedforms
360 are composed of sand size particles that were sorted by the wind and include sand dunes and ripples.
361 They are either well sorted by size or poorly sorted and typically cohesionless. Some of the ripples have
362 a slightly cohesive near surface layer (few kPa) a few centimeters thick (Sullivan et al., 2011). 2) Drift
363 deposits appear to be very fine grained dust (<10 μm) that has settled out of the atmosphere
364 (Christensen and Moore, 1992; Moore et al., 1999; Paton et al., 2016). This material is also effectively
365 cohesionless (and not load bearing). More cohesive soils have also been found. These soils have a
366 cohesive surface crust and/or break up into clods or blocks when deformed. Crusty and cloddy soils
367 have cohesions of less than 4 kPa and blocky soils have higher cohesions of 3-11 kPa (Moore et al., 1987;
368 Herkenhoff et al., 2008). Both are composed of dominantly sand size grains with some pebbles. The
369 cohesive soils in most cases are limited to surface layers of the order of centimeters thick and likely
370 formed by precipitation of salts from thin films of water interacting with the atmosphere (Haskin et al.,
371 2005; Tosca et al., 2004; Hurowitz et al., 2006; Martin-Torres et al., 2015).

372

373 **2.4 Internal Friction Angle**

374 The internal friction angle of sands depends on their grain size distribution, grain shape, particle
375 surface texture, grain arrangement and bulk density. Friction angles are determined by shearing
376 specimens under constant confining stress, by using either a direct shear box or a triaxial apparatus.
377 Shearing mobilizes irreversible volume changes. Loose sands decrease in volume due to the
378 entanglement of grains during shear; dense sands increase in volume due to disentanglement, providing
379 larger resistance to shear and higher friction angles. At the same density, angular particles provide
380 higher friction angles than rounded particles. As discussed above, the surficial Martian regolith at the
381 InSight landing site is expected to be composed of rounded particles in the range $\sim 150\text{-}250\ \mu\text{m}$ (fine
382 sand) (Golombek et al., 2008a, 2017). In this regard, shear tests carried out on lunar regoliths (Scott,
383 1987) or lunar regolith simulants (JSC-1 simulant or other crushed basalts, e.g., McKay et al., 1994;
384 Alshibli and Hasan, 2009; Vrettos, 2012) are not relevant, given the highly angular shape of their grains.

385 As shown in Delage et al. (2017), various Mars regolith simulants, that have been apparently selected
386 based on mineralogical considerations, are also somewhat angular. The Mojave Mars Simulant provided
387 by JPL (MMS, Peters et al., 2008) is crushed Miocene basalt, the Mars Soil Simulant-Dust provided by
388 DLR (MSS-D; Becker and Vrettos, 2016) is a 50/50 mix of crushed olivine and quartz sand (with a bimodal
389 grain size distribution curve and olivine particles finer than what is expected at the InSight landing site).
390 The Eifelsand simulant of DLR is a mix of crushed basalt and volcanic pumice sand (Delage et al., 2017).
391 In this respect, simulants based on quartz sands (e.g., WF34; Lichtenheldt, 2016) may be mechanically
392 more representative for what is expected to be present at the InSight landing site, as quartz sands show
393 mainly rounded to sub-angular grains.

394 Lee and Seed (1967) considered changes in friction angle with density in a terrestrial Sacramento
395 River (USA) sand, which is composed of rounded grains. These changes are compared in Figure 4 with
396 the friction angles of a Mojave simulant (a mix of MMS, containing alluvial sedimentary and igneous
397 grains from the Mojave Desert and basaltic pumice), MSS-D, and Eifelsand, determined with a direct
398 shear box at a bulk density of 1570 kg/m³ by Delage et al. (2017). The figure demonstrates the decrease
399 in friction angle at lower density with a good correspondence between the Sacramento River sand and
400 the Mojave simulant (angle of 38°, compared to 35° for MSS-D and 42° for Eifelsand, probably due to the
401 very angular and irregular shape of pumice particles). Extrapolation at bulk density of 1300 kg/m³
402 provides a friction angle between 28 and 30° for the surficial layer at the InSight landing site.

403 == Figure 4 about here ==

404 The changes in friction angle with depth can be estimated based on the changes in density shown in
405 Figure 3, assuming a density dependence of the friction angle ϕ corresponding to that of the Sacramento
406 River sand. A second order fit to the data results in

$$\phi = A\rho^2 + B\rho - C \quad (2)$$

407 where ρ is given in units of kg/m³. A , B , and C are constants with values of $-5.9772 \times 10^{-5} \text{m}^6/\text{kg}^2$,
408 $0.21583 \text{m}^3/\text{kg}$, and 152.88° , respectively. In the medium compacted case (Figure 3), the increase at 5
409 m is negligible, whereas the friction angle increases up to 36° in the very dense case. As commented
410 above, the increase in density and friction angle also involves the mobilization of dilating behavior of the
411 sand, which could have some consequence on the penetrability of the mole. Dilation mobilized during

412 penetration at the sand/mole interface results in an increase in radial stress that makes the penetration
413 less efficient, as a greater portion of the stroke energy is needed to mobilize the soil.

414 **2.5. Grain size Distribution**

415 We base our estimation of the average grain size distribution (GSD) within the InSight landing ellipse
416 using a combination of observations and modeling. We have previously used this approach to
417 extrapolate to the larger 10 cm particle size and hence determine the probability of obstruction of the
418 HP³ mole by a rock (Golombek et al., 2017). Here we extend the extrapolation down to the smaller 600
419 μm , an upper limit of the particles that may be present through eolian processes. The model
420 parameters are derived for the fragmentation that has produced the observable rocks through
421 meteorite impact, and therefore extrapolation into a size regime potentially dominated by eolian
422 processes has limited justification.

423 Our previous study applied the negative binomial (NB) fragmentation model (Charalambous,
424 2014/2015) to the rocks of the compiled HiRISE images from the InSight landing ellipse (Golombek et al.,
425 2017). We validated this approach by matching rock distributions from HiRISE images of Viking 2, Mars
426 Pathfinder, Spirit, and Phoenix to subsequent ground truth imaging. We predicted that the surface
427 population down to 10 cm is likely to be similar to that observed at Columbia Memorial Station (CMS)
428 (Golombek et al., 2017). The NB model is readily able to extrapolate the particle size distribution of a
429 surface population used to validate the model down to 5 cm in the case of Spirit and Phoenix.

430 In estimating a cumulative mass fraction of the regolith, it is necessary to match both the surface
431 rocks' size distribution, and the rock coverage expressed as a cumulative fractional area (CFA). To match
432 both in general requires an adjustment, in this case an addition, of material below the observable rock
433 size. The physical basis for such an addition is deposition of eolian material and subsequent mixing by
434 meteorite impact. This dilution of the fragmentation products by eolian material provides the observed
435 CFA. The eolian material can only be introduced for particle sizes below the saltation limit which we
436 take at the upper limit of 600 μm (Kok et al., 2012). Figure 5 shows the predicted grain size distribution
437 (GSD) based on these considerations down to the saltating upper size bound which, for the case of the
438 InSight landing site ellipse (E9), predicts the GSD $\sim 75\%$ by mass below 600 μm .

439 == Figure 5 about here ==

440 We can state that the GSD at the InSight landing site is likely to be close to the GSDs of the CMS and
 441 Phoenix landing sites, even though eolian processes might dominate at the InSight landing site. The
 442 thermal inertia in InSight landing ellipse has a value of about $200 \text{ J}/(\text{m}^2 \text{ K s}^{1/2})$, similar to that of CMS and
 443 Phoenix landing site. As the thermal inertia is dominated by particles of $100 \mu\text{m}$ or below in size, this
 444 suggests a common eolian component. On this basis, the predicted grain size distribution for the InSight
 445 landing site is expected to make a transition below $600 \mu\text{m}$ to match the observed GSD of the sand
 446 determined by the Phoenix microscope station (Pike et al., 2011).

447

448 **3. Regolith Thermophysical Properties**

449 This section compiles regolith material parameters needed to calculate subsurface temperatures at
 450 the InSight landing site. The energy balance of the shallow subsurface is governed by insolation,
 451 regolith thermal inertia, and heat diffusion into the deeper subsurface. The one dimensional heat
 452 diffusion equation can be written as:

$$\rho(z)c_p(T)\frac{\partial T}{\partial t} = \frac{\partial}{\partial z}k(z,P,\rho,T,\sigma)\frac{\partial T}{\partial z} \quad (3)$$

453 where ρ is density, c_p is specific heat, T is temperature, z is depth, P is CO_2 gas pressure, t is time, σ is
 454 ambient (overburden) pressure, the pressure exerted by the gravitational attraction of the mass of the
 455 column of regolith above the depth of interest, and k is thermal conductivity. Equation (3) is a second
 456 order differential equation, which can be solved by prescribing two boundary conditions: One is usually
 457 given by constant (or zero) heat flux at a depth, while the other is usually given in terms of the surface
 458 energy balance. For periodic insolation forcing, the surface energy balance takes the convenient form

$$\sigma_B \varepsilon T^4 = (1 - A)S + \varepsilon R + I \sqrt{\frac{\pi}{p}} \left. \frac{\partial T}{\partial z'} \right|_{z'=0} \quad (4)$$

459 where σ_B is the Stefan-Boltzmann constant, ε is surface emissivity, A is albedo, S is total solar radiative
 460 flux including scattered radiation, R is the thermal radiative flux from the atmosphere towards the
 461 surface, p is the period of the forcing, and $z' = z/d_e$ is depth normalized to the thermal skin depth $d_e =$

462 $\sqrt{k\rho/\rho c_p\pi}$. In Equation (4), all material parameters have been absorbed in the thermal inertia I , which
463 is defined as

$$I = \sqrt{k\rho c_p} \quad (5)$$

464 Equation (5) is only valid when thermal conductivity is constant, which is not the case (see below).
465 However, constant thermal inertia is a convenient way to describe the response of surface temperatures
466 to insolation changes, and it is thus a widely used approximation. However, care must be taken when
467 converting thermal inertia to material parameters like thermal conductivity, since different
468 combinations of material parameters govern the temperature at the surface (thermal inertia) and in the
469 subsurface (thermal diffusivity, see below). The expected values of material parameters and their
470 dependencies will be discussed for the InSight landing site below.

471 **3.1. Surface Emissivity**

472 Emissivity ε is defined as the ratio of emitted specific radiance I_r ($\text{W}/(\mu\text{m m}^2 \text{ sr})$) to the black-body
473 radiance B of a surface at temperature T . Emissivity is a function of the wavelength λ and viewing
474 angle, but the angle dependence is commonly assumed to be negligible and the radiative heat flux
475 density q_{rad} (W/m^2) of thermal emission can be represented via hemispherical integration as

$$q_{rad} = \pi \int_0^\infty \varepsilon(\lambda) B(T, \lambda) d\lambda \quad (6)$$

476 Often, ε is assumed to be a constant, i.e., $\varepsilon = \varepsilon_q$, where ε_q is the weighted spectral average
477 emissivity. Equation 6 can then be reduced to a form similar to the Stefan-Boltzmann Law:

$$q_{rad} = \varepsilon_q \sigma_B T^4. \quad (7)$$

478 where σ_B is the Stefan-Boltzmann constant. This approximation is usually sufficient for thermal models
479 but has a systematic error as a function of T if ε varies with wavelength.

480 Instruments for Mars surface thermal emission observations include the Thermal Emission
481 Spectrometer (TES) on Mars Global Surveyor (Christensen et al., 2001), the Thermal Emission Imaging
482 System (THEMIS) on Mars Odyssey (Christensen et al., 2003a), the Mini-Thermal Emission Spectrometer
483 (Mini-TES) on the Mars Exploration Rovers (Christensen et al., 2004a, b), the Planetary Fourier

484 Spectrometer (PFS) on Mars Express (Formisano et al., 2005) and the Ground Temperature Sensor of the
485 Rover Environmental Monitoring Station (REMS – GTS) on the Mars Science laboratory (Gomez-Elvira et
486 al., 2012). It should be noted that interpretation of thermal emission is ambiguous because two
487 unknowns, i.e., surface temperature and emissivity, contribute to the radiance, while only a single
488 quantity is measured. Therefore, observations aim at measuring radiance close to the Christiansen
489 wavelength, the wavelength at which the real part of silicate particle refractive index matches that of
490 the atmosphere, and emissivity is close to unity (Conel, 1969).

491 Assuming soil physical and compositional properties similar to those observed at the two Mars
492 Exploration Rovers landing sites (Golombek et al., 2005, 2008a; Yen et al., 2005), the InSight site is
493 expected to be covered by basaltic sand, possibly covered in places with a fine, higher albedo dust. We
494 use Mini-TES spectra analyzed by Ruff et al. (2006) as a basis for emissivity estimates. These spectra are
495 shown in Figure 6. They correspond to a bright dust drift (green), a basalt rock cleaned of dust by the
496 Rock Abrasion Tool (blue), and to the darker sand exposed at surfaces disturbed by the rovers at Gusev
497 crater (red) and Meridiani Planum (black). Data affected by the set of strong CO₂ absorption lines near
498 15 μm wavelength have been removed.

499 == Figure 6 about here ==

500 The constant emissivity ϵ_q that best represents the heat flux from the surface is a function of
501 composition and surface temperature, because the peak of the blackbody emission changes significantly
502 within the range of expected temperatures. For the dark soil, the expected value for ϵ_q is in the range of
503 0.97 to 0.985, with less than 0.5 % change with temperature. The bright dust and basalt have a similar
504 ϵ_q of 0.96 at 285 K, which increase by 2 % and decrease by 1.5 % towards 185 K, respectively.
505 Therefore, based on remote sensing and in-situ data, a constant emissivity value of 0.98 (+1% /-2%) is
506 suitable for both thermal modeling and surface temperature derivation at the InSight landing site, and
507 the stated uncertainty is equivalent to a deviation in derived thermal inertia of <20 J/(m² K s^{1/2}) in the
508 model of Vasavada et al. (2017). Examples of weighted average thermal emissivities for the HP³
509 radiometer filters are given in Table 2.

510 == Table 2 about here ==

511

512 3.2. Surface Thermal Inertia

513 Thermal inertia describes the resistance to a change in temperature of the upper 2–30 cm of the
514 surface. Fine particles change temperature quickly and therefore have low thermal inertia; higher
515 thermal inertia surfaces are composed of sand, duricrust, rock fragments, or a combination of these
516 materials. Bulk orbital thermal inertia observations of Mars include values derived from: (1) Viking
517 Infrared Thermal Mapper (IRTM) data at ~60 km per pixel (Kieffer et al., 1977; Palluconi and Kieffer,
518 1981), (2) Mars Global Surveyor TES data (Christensen et al., 1992) at 8 pixels per degree (Mellon et al.,
519 2000; Christensen et al., 2001) and at 20 pixels per degree (Putzig et al., 2005; Putzig and Mellon, 2007),
520 and (3) Mars Odyssey THEMIS data at ~100 m/pixel (Christensen et al., 2004c; Fergason et al., 2006a;
521 Fergason et al., 2012). Surface thermal inertia measurements were also obtained by the Miniature
522 Thermal Emission Spectrometer (Mini-TES) on the Spirit and Opportunity rovers during their traverses
523 (Christensen et al., 2003b; Fergason et al., 2006b). In addition, Curiosity determined thermal inertia
524 from Ground Temperature Sensor (GTS) measurements from the Rover Environmental Monitoring
525 Station (REMS) instruments (Hamilton et al., 2014, Vasavada et al., 2017).

526 Bulk thermal conductivity ranges over 3 orders of magnitudes on Mars as a function of the physical
527 state of the (sub-)surface (compared to small factors for ρ and c_p as a function of the porosity,
528 temperature, composition, etc., compare Equation (5)). I is virtually independent of the product ρc_p ,
529 whose value is generally close to $\sim 10^6$ J/(m³ K) (Neugebauer et al., 1969; Fergason et al., 2006a), and is
530 mainly controlled by k . More precisely,

$$k \approx \frac{I^2}{8 \cdot 10^5} \text{ W/(m K)} \quad (8)$$

531 for temperatures and surface densities at the InSight landing site. On Mars, thermal inertia values have
532 largely been derived from remote measurements. Because of the strong dependence of its value on
533 grain size and degree of cementation, Putzig (2006) distinguished between dust (28-135 J/(m² K s^½)),
534 sand (135-630 J/(m² K s^½)) and duricrust (252-513 J/(m² K s^½)). Paton et al. (2016) gave a value for I of 81
535 to 125 J/(m² K s^½) for dust around the Viking 1 footpads from direct measurements.

536 The highest resolution TES nighttime thermal inertia determination of the InSight landing site (Putzig
537 and Mellon (2007) at 20 pixels per degree range from 138 to 284 J/(m² K s^½) and average 218 J/(m² K s^½)
538 (n=314). A regional thermal inertia map (100 m spatial scale) was generated for the landing site
539 (Golombek et al., 2017) from predawn temperature data acquired by THEMIS band 9 (12.57 μ m)

540 (Christensen et al., 2004c) between Mars Year 30 and 32 during low-dust seasons to minimize the
541 atmospheric impact on the derived values. The resulting thermal inertia map displays values ranging
542 from $\sim 70 \text{ J}/(\text{m}^2 \text{ K s}^{1/2})$ to $390 \text{ J}/(\text{m}^2 \text{ K s}^{1/2})$, but 99% of the area has a thermal inertia of 130 to $220 \text{ J}/(\text{m}^2 \text{ K s}^{1/2})$.
543 Within the landing ellipse, the range is even smaller, demonstrating high thermophysical
544 homogeneity at the 100 m scale over the entire landing region. The median regional thermal inertia is
545 $\sim 180 \text{ J}/(\text{m}^2 \text{ K s}^{1/2})$, corresponding to cohesionless $\sim 170 \mu\text{m}$ material (fine sand) based on laboratory work
546 and theoretical relationships (Presley and Christensen, 1997a; Piqueux and Christensen, 2011). Higher
547 thermal inertia values are expected to be associated with medium to coarse sand, and will likely include
548 mixtures of grain sizes, including larger clasts such as those surfaces observed at Gusev crater
549 (Golombek et al., 2005, 2008a; Fergason et al., 2006b). The corresponding diurnal skin depth values
550 (i.e., depth at which maximum amplitude is attenuated to 37% of its surface amplitude) is a maximum of
551 $\leq 6 \text{ cm}$, indicating that the upper few cm of the surface layer are characterized by these thermal inertia
552 values. The lack of seasonal variations in thermal inertia indicates that the same thermal inertia and
553 materials extend a few tens of cm below the surface (Golombek et al., 2017).

554 The lowest thermal inertia values in the landing region (e.g., $\sim 70 \text{ J}/(\text{m}^2 \text{ K s}^{1/2})$) are rare, and typically
555 are observed within depressions probably that trap atmospheric dust and very fine sand, or on the lee
556 side of positive topographic features (Golombek et al., 2017). These low inertia values could result from
557 fine sand ($100\text{--}200 \mu\text{m}$) with a very thin coating ($<1\text{--}2 \text{ mm}$) of dust (several μm diameter particles). The
558 highest thermal inertia values (i.e., $350\text{--}390 \text{ J}/(\text{m}^2 \text{ K s}^{1/2})$) are also uncommon, associated with crater
559 rims and ejecta blankets, as expected for rocky ejecta craters, but not bedrock at the 100 m spatial
560 scale. Regolith induration is not inconsistent with the derived thermal inertia values, however thermal
561 modeling of cemented regolith shows that the volume of the cementing phase would need to be
562 minimal (e.g., typically $<0.1\%$ in volume) with little impact on the mechanical properties (Piqueux and
563 Christensen, 2009a). Comparison of the cohesion of surface soils at other landing sites with their
564 thermal inertia would limit the cohesion to less than a few of kPa, consistent with very weakly bonded
565 soils on Earth (Golombek et al., 1997, 2008a).

566

567 **3.3. Surface Albedo**

568 The albedo, or surface reflectivity or brightness of reflected solar energy from the surface in which
569 the viewing geometry has been taken into account, has been measured globally by both IRTM and TES at

570 1 pixel and 8 pixels per degree, respectively (e.g., Pleskot and Miner, 1981; Christensen et al., 2001).
571 The albedo can, for example, be used to infer the dustiness of the surface, as very dusty areas exhibit
572 very high albedo (and, in addition, very low-thermal inertia) (Christensen and Moore, 1992; Moore and
573 Jakosky, 1989; Mellon et al., 2008; Putzig et al., 2005; Golombek et al., 2008a). The amount of dust
574 cover at the landing sites was also evaluated using the TES dust cover index (16 pixels per degree), which
575 includes a more explicit measure of the particle size and the amount of dust coating the surface (Ruff
576 and Christensen, 2002).

577 The albedo of the InSight landing site is about 0.25 from IRTM (Pleskot and Miner, 1981) and 0.24
578 from TES (Christensen et al., 2001). This relatively high albedo is consistent with atmospherically
579 deposited dust, which is consistent with its relatively high dust cover index (Ruff and Christensen, 2002).
580 However, thermal inertia values are nowhere dominated by very fine material at the 100 m scale
581 suggesting that dust may form an optically thick but thermally thin coating (hundreds of μm) on most
582 surface materials in this region of Mars. This interpretation is supported by the similarity of the dust
583 cover index in the InSight landing site region (0.94) with the Viking Lander 2 site and dusty locations of
584 the Gusev cratered plains explored by Spirit (e.g., Golombek et al., 2005, 2006), both of which had very
585 thin dust coatings.

586 All previous landers on Mars have modified the surface during landing (e.g., Moore et al., 1987;
587 Golombek et al., 1999; Squyres et al., 2004; Soderblom et al., 2004; Smith et al., 2009; Arvidson et al.,
588 2014; Daubar and McEwan, 2015). The InSight lander will use pulsed retropropulsive thrusters to slow
589 itself during landing. The thrusters on InSight are the same as those used by the Phoenix lander, which
590 dispersed 5-18 cm of soil exposing water ice when landing (Mehta et al., 2011). Modeling of this process
591 showed that pulsed thrusters lead to explosive erosion via cyclic shock waves that fluidize soils,
592 producing ten times greater erosion than conventional jets (Mehta et al., 2011, 2013). Consideration of
593 these effects for InSight landing indicates that generally circular depressions will form at the jet
594 impingement locations, but they will not be large enough to appreciably alter the surface topography at
595 the lander footpad locations and thus won't pose a risk to landing safely (Golombek et al., 2017).
596 Nevertheless, surface soils will be dispersed away from the lander with sand and pebbles being eroded
597 from the jet impingement locations and deposited away from the spacecraft.

598 The thin coating of fine-grained dust present at the landing site will be dispersed into the
599 atmosphere at the time of landing, reducing the albedo of the surface around the lander. This has been

600 observed to occur around previous landers, and in the cases of Phoenix and Mars Science Laboratory
601 (Curiosity) the effect can be measured using relative albedo measurements in HiRISE images (Daubar
602 and McEwan, 2015). The quantity of albedo change and rate of subsequent brightening varied
603 depending on the particular piece of hardware; for the MSL descent stage, the albedo was initially
604 lowered by ~50%. After the initial darkening, images show a rapid initial brightening that slowed over
605 time, following a logarithmic function. The majority of the blast zone faded to ~90% of the initial albedo
606 by ~500 days after landing, but the darkest areas have not faded completely. Although it is located at
607 high latitudes, the Phoenix landing site is in some ways a better analogy for InSight due to the same
608 landing thrusters; however, monitoring of the Phoenix site is complicated by seasonal activity and
609 limitations to orbital observations. The Phoenix landing reduced the surroundings to ~60-80% of the
610 pre-landing albedo. Before subsequent orbital images could be taken in the same season, the blast zone
611 disappeared, presumably due to seasonal frosts redistributing surface dust.

612 Based on these observations and the relatively dusty nature of western Elysium Planitia, we would
613 expect similar changes to the InSight landing site, where the surface albedo can be expected to be
614 reduced by ~20-50% upon landing, then exhibit a rapid initial brightening, and then gradually return to
615 the surrounding albedo over the next several Mars years. The reduction in albedo will warm the surface
616 and the deposition of sand and pebbles from the thrusters could also have a thermal effect.

617

618 **3.4. Thermal Conductivity**

619 This section describes recommended values for the thermal conductivity k of the regolith expected at
620 the InSight landing site, based on orbital data and published laboratory/theoretical work. Unless
621 otherwise specified, the regolith is treated as an idealized discontinuous medium composed of spherical
622 basaltic grains in stagnant CO₂ gas. The relationship between bulk regolith conductivity and various
623 controlling factors (i.e., pressure, temperature, grain size, porosity, etc.) is quantitatively described in
624 the literature for a wide range of planetary configurations of atmospheric pressures, compositions,
625 regolith properties, etc. For the specific case of the InSight landing region, these relationships have
626 been tailored to the expected subsurface properties for simplicity, and are presented here. We will first
627 discuss an appropriate choice for the simple case of constant thermal conductivity and then present the
628 more general case of temperature and depth dependency.

629 Thermophysical properties of the landing region have been characterized from orbital data acquired
 630 by the Thermal Emission Imaging System (THEMIS) (Christensen et al., 2004c). In the landing ellipse,
 631 thermal inertia I values derived from temperature measurements typically range from 130 to 220 J/(m² K
 632 s^{1/2}) with a median value of ~180 J/(m² K s^{1/2}) (Golombek et al., 2017). Using relationships established in
 633 the laboratory (Presley and Christensen, 1997b) the expected regolith thermal conductivity is $0.017 < k <$
 634 0.048 W/(m K) with median value of 0.032 W/(m K) corresponding to ~150-170 μ m unconsolidated
 635 grains (Golombek et al., 2017).

636 Published thermophysical studies of Martian subsurface temperatures generally use fixed I or k (as
 637 opposed to temperature or pressure-dependent values), because these dependencies are not
 638 straightforward to determine, and because they result in small overall conductivity (Piqueux and
 639 Christensen, 2011) and surface temperature (Kieffer, 2013) changes at the expense of longer processing
 640 time. In the context of the InSight heat-flow experiments. However, subtle conductivity variations may
 641 need to be accounted for. Therefore, the dependence of thermal conductivity on gas pressure,
 642 porosity/density, temperature and overburden pressure/stress will be considered in this section.

643 Because of the discontinuous nature of the solid phase, with inter-grain regions impeding the flow of
 644 heat from grain to grain, the bulk regolith conductivity is strongly influenced by the pore-filling CO₂ gas
 645 conductivity (~0.01 W/(m K) at 220K). In rarefied gas environments, where the mean free path of gas
 646 molecules is similar to the volume that encapsulates them (i.e., the pore space) as is the case in the
 647 Martian regolith, small pressure variations can result in noticeable bulk conductivity changes.
 648 Laboratory experiments have quantified this effect (Fountain and West, 1970; Presley and Christensen,
 649 1997a), and numerical models also include it (Piqueux and Christensen, 2009b). The effect of gas
 650 pressure on the bulk conductivity is described by the empirical Equation 9 (modified from Presley and
 651 Christensen, 1997b):

$$k = (CP^{0.64})d^{-0.125 \log(\frac{P}{K})} \quad (9)$$

652 with $C = 0.001262$, $K = 107990$ hPa, d is the grain diameter in μ m, and P is pressure in hPa. This
 653 equation is dimensionally unbalanced and was derived by Presley and Christensen (1979b) from log-log
 654 plots of laboratory measurements of thermal conductivity as a function of gas pressure for different
 655 grain sizes in the range of 11 to 900 μ m. The equation is not based on a theoretical analysis of heat
 656 transfer in granular media. Figure 7 shows the predicted variation of the bulk conductivity as a function
 657 of the atmospheric pressure using Equation (9). For a given location, the ~30% seasonal variation of the

658 atmospheric pressure due to the CO₂ cycle at the poles (Leighton and Murray, 1966; Hess et al., 1979)
659 induces ~10% of conductivity variation. A simplification of Equation (9) gives (Figure 10):

$$k(P + \Delta P) = k_0(P) \cdot (1 + A \cdot \Delta P + B \cdot \Delta P^2) \quad (10)$$

660 where $k(P + \Delta P)$ is the thermal conductivity at a pressure with ΔP the atmospheric pressure deviation
661 (in hPa) from the local mean pressure P and $k_0(P)$ the nominal regolith conductivity at pressure P . $A =$
662 5.173 hPa^{-1} and $B = -2.416 \cdot 10^{-1} \text{ hPa}^{-2}$ are coefficients derived from a fit based of Equation (10) and Figure
663 7. Coefficients in Equations 9 and 10 are only valid for the range of range of grain sizes and pressures
664 used in the Presley and Christensen (1979b) laboratory experiments.

665 == Figure 7 about here ==

666 In addition, we note that Equations 9 and 10 do not apply for strongly cemented material. With
667 indurated material, the relatively low pore-filling gas conductivity that enables heat transfer in the high
668 impedance inter-grain region is replaced by high-conductivity inter-granular material (solids such as salts
669 or ices are several orders of magnitude more conductive than rarefied CO₂ gas) and control the
670 dependence of k on the temperature and pressure (Piqueux and Christensen, 2009b). As a result, the
671 bulk thermal conductivity of cemented regolith is less dependent on atmospheric pressure variations.
672 Equation 10 only provides an upper limit to the dependence on pressure. We note that the
673 interpretation of remote sensing thermal infrared data is not consistent with a fully encrusted regolith,
674 but does not exclude a very slight surface induration (Golombek et al. 2017). We anticipate Equation 10
675 to be adequate in the nominal landing region.

676

677 Laboratory experiments (Fountain and West, 1970; Presley and Christensen, 1997a) and theoretical
678 considerations (Piqueux and Christensen, 2009b) indicate that the porosity of the Martian regolith
679 partially controls the bulk thermal conductivity. High porosities are generally associated with lower bulk
680 conductivities.

681 Fountain and West (1970) (their Figure 3) used samples typically finer than those expected at the
682 InSight landing site (i.e., 37-62 μm), and they found an ~200+% increase in bulk conductivity for a ~50%
683 increase of the density (ignoring their very low density samples). Based solely on numerical modeling,
684 Piqueux and Christensen (2009b) found a doubling of the bulk conductivity associated with a doubling of

685 the density (their Figure 7). Presley and Christensen (1997b) observed a ~30% increase of the bulk
686 conductivity for a 30% increase of the density for Kyanite samples at all pressures tested, a trend
687 consistent with modeling by Piqueux and Christensen (2009b), but significantly less pronounced than
688 that by Fountain and West (1970). We propose to adopt a linear conductivity dependency on density
689 that conforms with the most recent laboratory work models (i.e., work by Presley and Christensen
690 (1997b), and Plesa et al. (2016)):

$$k(\rho+\Delta\rho)=k_0(\rho)\cdot(1+0.005\cdot\Delta\rho) \quad (11)$$

691 where $k(\rho+\Delta\rho)$ is the thermal conductivity with $\Delta\rho$ the change in regolith density (in %) from the
692 nominal density ρ , and $k_0(\rho)$ the conductivity with the nominal density.

693 Under most Martian surface conditions, including those expected at the InSight landing site, although
694 radiative heat transfer probably dominates in the atmosphere (e.g., Martinez et al., 2014) it is small
695 compared to other mechanisms in the regolith (Vasavada et al., 1999) and is therefore ignored in the
696 analysis here. Apart from radiative heat transport, temperature also controls the pore-filling gas
697 conductivity, as well as the solid phase conductivity. The solid phase conductivity is only weakly linked
698 to the bulk regolith conductivity, such that temperature induced variations of the solid phase
699 conductivity can usually be ignored.

700 A theoretical quantification of the bulk conductivity dependency on the gas conductivity is a difficult
701 problem because of the complex geometry of the gaseous phase and its relationship to the solid phase.
702 Increasing the regolith temperature increases the intrinsic conductivity of the pore filling gas (Vesovic et
703 al., 1990), but also decreases the mean free path, reducing the efficiency of the gaseous heat transfer. A
704 quantitative comparison of these two opposite mechanisms requires numerical modeling and indicates
705 that the reduction of the mean free path has a very small effect compared to the general bulk gas
706 conductivity increase with temperature (Piqueux and Christensen, 2009b; 2011). As a result, increasing
707 the temperature in stagnant CO₂ gas and with pressures consistent with Mars increases the bulk
708 conductivity of the regolith, as confirmed by laboratory measurements (Fountain and West, 1970).
709 Piqueux and Christensen (2011) compared the temperature effect on k predicted by their model with
710 the data published by Fountain and West (1970), and results are shown in Figure 8.

711 == Figure 8 about here ==

712 Generally, the numerical model predicts a larger temperature-dependency than observed in the
713 laboratory, over a wide range of material density and temperatures. While Fountain and West (1970)
714 do not formally provide a relationship between temperature and bulk conductivity, their data indicates
715 a ~15-20% increase in bulk conductivity over 100K (Figure 8), in line with the expected increase in pore-
716 filling gas conductivity over this range of temperatures. For comparison, a Piqueux and Christensen
717 (2011) model emulating these laboratory conditions found a ~30% increase over 100K (Figure 8), which
718 is remarkably close to the experimental observations given the numerous modeling assumptions. Given
719 that the temperature dependence of the pore fill gas is the major contribution to the thermal change,
720 we propose as square-root dependence of regolith thermal conductivity on temperature, consistent
721 with the kinetic theory of gases. Bulk conductivity as a function of temperature $k(T)$ where T is
722 temperature (in K) is then given by

$$k(T) = k_0(T_0) \sqrt{T/T_0} \quad (12)$$

723 where T_0 and k_0 are the nominal temperature (in K) and regolith conductivity (in W/m K), respectively. A
724 fit to the data by Fountain and West (1970) is shown in Figure 8, demonstrating that this approximation
725 is appropriate for the range of temperatures expected to be encountered on Mars. Again, this trend
726 only applies for unconsolidated material in the presence of rarefied gas. In the case of a duricrust,
727 Equation 12 does not apply because the gas conductivity does not dominate the bulk conductivity, and
728 the thermal conductivities of solid (cementing) phases generally decrease with increasing temperature,
729 following a trend opposite to Equation 12. As a result, the dependence of k with T in the case of
730 indurated material is nonlinear and too complicated to predict without ad hoc models (Piqueux and
731 Christensen 2011).

732

733 An increase of the confining pressure, for example as a result of the progression of the HP³ mole, is
734 expected to result in an increase of the bulk regolith conductivity by increasing the contact area
735 between grains (Hertz, 1895), hence facilitating the flow of heat from grain to grain at the expense of
736 the relatively inefficient (but dominating) gaseous heat transfer. Elasticity theory suggests that contact
737 area, and thus thermal conductivity, should scale with stress σ to the power of one third, but different
738 scaling relations with other power law dependence have also been suggested (e.g., Pilbeam and Vaisnys,

739 1973). However, current laboratory data is most consistent with a power law scaling close to the
 740 classical value, and

$$k = k_0 \left(\frac{\sigma}{\sigma_0} \right)^{1/3} \quad (13)$$

741 has been established for monodispersed spheres as well as for lunar analogue material (Sakatani et al.,
 742 2016), where k_0 is the conductivity at pressure σ_0 .

743 Apart from the action of the HP³ mole, stress anisotropy of the regolith itself could have an influence
 744 on regolith thermal conductivity. Stress anisotropy is generally described in terms of the dimensionless
 745 coefficient of lateral stress

$$K_0 = \frac{\sigma_h}{\sigma_v} \quad (14)$$

746 where σ_h and σ_v are the stresses in the horizontal and vertical directions, respectively. For normally
 747 consolidated soils, K_0 is usually between 0.4 and 0.5, consistent with Jaky's formula $K_0 = 1 - \sin(\phi)$ (Jaky,
 748 1944) for angles of internal friction ϕ close to 30°. Stress anisotropy may then introduce anisotropy into
 749 the thermal conductivity, i.e., conductivity may vary between the horizontal and vertical directions.
 750 While this effect may be pronounced on airless bodies, it will be largely mitigated on Mars by the pore
 751 filling CO₂ gas.

752 In order to estimate the magnitude of the expected effect, the contribution of the pore filling gas to
 753 the total thermal conductivity can be estimated by writing $k_{h,v} = k_{sol,h,v} + k_{gas}$, where k_{sol} and k_{gas} are the
 754 solid and gas conductivity part of the thermal conductivity, and subscripts h and v refer to the horizontal
 755 and vertical direction, respectively. Using Equations (13) and (14), thermal conductivity in the horizontal
 756 direction can then be expressed as

$$k_h = k_{sol,v} K_0^{1/3} + k_{gas} \quad (15)$$

757 Hütter et al. (2008) give thermal conductivities of 0.008 and 0.057 W/(m K) for 100 – 200 μm
 758 diameter glass beads under vacuum and 8 hPa pressure conditions, respectively, and we therefore
 759 assume $k_{gas} = 0.049$ W/m K) and $k_{sol,v} = 0.008$ W/(m K) respectively. Note that these grain sizes closely
 760 correspond to the expected grain size range at the InSight landing site derived from surface thermal
 761 inertia, which results in 150 μm diameter grains. Then, for normally consolidated soil, $K_0 = 0.5$ and k_h is

762 expected to be smaller than k_v by about 2-3%. Note that this effect is even less pronounced for larger
763 grain sizes, and can likely be ignored in the context of the InSight mission.

764

765 3.5. Specific Heat

766 The specific heat of rocks and soils at low temperatures has been studied for lunar samples (Robie et
767 al., 1970; Fujii and Osako, 1973; Hemingway et al., 1973), and a strong temperature dependence has
768 been found. The suite of materials studied includes particulate material such as lunar fines and soils,
769 but brecciated lunar rocks as well as basalts have also been studied. A best fit to the lunar soils data was
770 given by Hemingway et al. (1973) and the specific heat can be approximated as

$$c_p = -A + BT + CT^2 - DT^3 + ET^4 \quad (16)$$

771 where c_p is specific heat in units of J/(kg K), and A , B , C , D , and E are constants with values 23.173 J/(kg
772 K), 2.127 J/(kg K²), 1.5008×10^{-2} J/(kg K³), 7.3699×10^{-5} J/(kg K⁴), and 9.6552×10^{-8} J/(kg K⁵), respectively,
773 and T is temperature in K. This best fitting formula is accurate to within 2 percent down to 200 K and to
774 within 6% down to 90 K. The fit is shown along with the data in Figure 9.

775 == Figure 9 about here ==

776 Measurements on lunar material are in good agreement with a thermophysical model of Winter and
777 Saari (1969), measurements on the physical properties of meteorites performed by Yomogida and
778 Matsui (1983), and meteorite specific heat measurements by Consolmagno et al. (2013). It may be
779 worth noting that a trend exists with respect to the iron content of the samples, with low iron
780 corresponding to high specific heat (Yomogida and Matsui, 1983). The contribution of the gas phase to
781 the bulk specific heat of a soil is negligible when compared to the solid phase and is usually ignored
782 (Piqueux and Christensen, 2011).

783 While specific heat thus shows a strong temperature dependence, this is only relevant if the near
784 surface regolith layer is considered. At depths below a few tens of cm, near surface temperature
785 perturbations rapidly decay (e.g., Grott et al., 2007; Kieffer, 2013) such that the regolith can be assumed
786 isothermal for the purpose of determining its specific heat. For the InSight landing site, average regolith
787 temperatures vary between 220 and 240 K (Plesa et al., 2016), corresponding to specific heat values of
788 612 and 653 J/(kg K) such that $c_p = 630$ J/(kg K) may be assumed.

789

790 3.6. Thermal Diffusivity

791 Thermal conductivity and specific heat are the most useful quantities in terms of modeling thermal
792 fluxes in the regolith and are probably the most physically meaningful. In practical applications,
793 however, they are often replaced by derived quantities that are either directly measurable or
794 convenient shorthand in equations. Apart from thermal inertia, which describes the reaction of surface
795 temperatures to harmonic temperature forcing and was introduced in section 3.2, thermal diffusivity
796 can be used to describe heat diffusion in the subsurface. Thermal diffusivity κ is defined as

$$\kappa = k / (\rho c_p) \quad (17)$$

797 where k is thermal conductivity, ρ is density, and c_p is specific heat. It is a particularly useful quantity if
798 material parameters can be assumed to be constant, and in this case the heat diffusion equation
799 (Equation 3) takes a particularly convenient form. As can be seen from Equation (17), an increase in
800 thermal conductivity has the effect of a corresponding decrease in specific heat, which implies that
801 thermal diffusivity is somewhat less sensitive to changes in density (which is most sensitive to porosity in
802 the regolith) than thermal conductivity. Over a narrow temperature and depth range, κ can therefore
803 be approximated as a constant, thus facilitating analytical solutions of the heat conduction equation. It
804 is worth noting that estimates of thermal diffusivity from the attenuation of the diurnal temperature
805 wave on the Moon did not show any systematic effects below a depth of 50 cm (Langseth et al., 1976),
806 and this may be a valid approximation for the Martian subsurface as well. In this case, $\kappa = 3.6 \times 10^{-8} \text{ m}^2/\text{s}$
807 would be a reasonable estimate at the InSight landing site.

808 For planetary regoliths in general, it is the thermal conductivity whose effect dominates the behavior
809 of κ which on Mars can span two orders of magnitude and be strongly temperature-dependent, whereas
810 the range of both density ρ and specific heat c_p are usually rather narrowly constrained. If depth
811 dependence of thermal diffusivity is deemed to be important, appropriate values for $\kappa(P, \rho, T, c_p(T))$ can
812 easily be computed by inserting Equations 10, 11, 12, and 16 into Equation 17.

813

814 4. Regolith Elastic Properties

815 This section deals with the elastic properties of the regolith, which characterize its influence on the
816 seismic wavefield as recorded by the SEIS (Seismic Experiment for Interior Structure) instrument. The
817 relevant parameters discussed here are compressional wave velocity v_p , shear wave velocity v_s ,
818 Poisson's ratio ν which can be derived from these velocities, elastic modulus E which can be expressed
819 in terms of the above quantities and density ρ , as well as the seismic quality factor Q .

820 4.1. Seismic Velocities and Poisson's ratio

821 Poisson's ratio ν describes the relation between transverse strain ε_{\perp} and axial strain ε_{\parallel} when a
822 uniaxial stress is applied

$$\nu = - \frac{d\varepsilon_{\perp}}{d\varepsilon_{\parallel}} \quad (18)$$

823 It is directly related to the seismic P- and S-wave velocities v_p and v_s by

$$\nu = \frac{\left(\frac{v_p}{v_s}\right)^2 - 2}{2 \left(\left(\frac{v_p}{v_s}\right)^2 - 1\right)} \quad (19)$$

824 with higher values of ν related to smaller shear resistance, and higher v_p / v_s .

825 In contrast to thermophysical properties, for which estimates can be based on remote sensing data
826 from Mars, or other mechanical properties, for which data are available from other Martian landing
827 sites, there are currently no in situ measurements of seismic velocities of the Martian regolith.
828 Estimates thus have to be based on laboratory experiments with analogue materials on Earth while also
829 considering field and lab data gathered for lunar regolith and terrestrial sands.

830 Both v_p and v_s were determined by Delage et al. (2017) for three Martian regolith soil simulants
831 under various confining pressures corresponding to lithostatic stresses from 5 m to more than 60 m
832 depth on Mars. The Mojave simulant, provided by JPL, is a mixture of MMS simulant, containing alluvial
833 sedimentary and igneous grains from the Mojave Desert, with basaltic pumice. The Eifelsand simulant
834 from DLR is a mixture of crushed basalt and volcanic pumice sand. The MSS-D simulant, also from DLR,
835 is an artificial sediment made of a 50/50 mixture of crushed olivine and quartz sand, with a bimodal
836 grain-size distribution, and olivine particles smaller than expected at the InSight landing site. As the

837 MSS-D particles are in the silt-size range (50% of particles smaller than 70 μm , and as small as 2 μm),
838 much finer than the particle sizes estimated for the regolith at the landing site (Golombek et al., 2017),
839 and are angular rather than rounded, the results more relevant to the InSight landing site are those for
840 the Mojave and Eifelsand simulants. The ejecta that form the Martian regolith are expected to be
841 rounded due to long term exposure to wind action in low atmospheric pressure conditions, in contrast
842 to lunar regolith particles that are not submitted to any wind and, as a result, are more angular. The
843 Mojave simulant contains both rounded and more angular grains and their particle size distribution is
844 closer to the landing site estimates, at least when using only particles smaller than 2 mm, as was done in
845 the laboratory measurements.

846 During the laboratory tests on Mojave simulant, Delage et al. (2017) observed no effect of stress
847 cycles on the values of seismic velocities, and hence no difference between the effect of either plastic
848 (first stress application) or elastic response along the compression strain. They found that the increase
849 in velocity was more sensitive to the increase in inter-granular forces resulting from an increase in
850 confining stress, and, to a lesser extent, to the corresponding increase in density. Data along three
851 successive stress paths as well as from tests carried out on two different samples showed good
852 agreement. The smallest confining stress used in these tests was 25 kPa, which approximately
853 corresponds to 5 m depth on Mars, so the properties of the regolith at shallower depth have to be
854 extrapolated.

855 For all regolith simulants, a power-law increase of velocities with depth was observed, defined in
856 relation to confining stress σ'_0 (in kPa) by an empirical law (Santamarina et al., 2001) given as:

$$v = \alpha \left(\frac{\sigma'_0}{1 \text{ kPa}} \right)^\beta \quad (20)$$

857 and where α and β are experimentally determined. α is the velocity of the velocity of the material
858 subjected to 1 kPa confinement; β is non-dimensional. This kind of velocity-depth dependence is also
859 common for terrestrial soils (e.g., Faust, 1951; Prasad et al., 2004). Fitting the laboratory measurements
860 for v_p resulted in a value of 0.3 for the exponent β and, using the velocity values of 250 m/s at 25 kPa
861 and 600 m/s at 500 kPa, a value of $\alpha = 95$ m/s for the compressional velocity at 1 kPa confining stress.
862 Surface velocities are derived assuming an atmospheric pressure of 0.6 kPa, and 81.5 m/s and 48.8 m/s
863 for P- and S-waves, respectively. Theoretical estimates (Santamarina et al., 2001) based on contact
864 theory result in values of 1/6 for β for Hertzian contacts between elastic spheres and 1/4 for cone to

865 plane contacts (expected for rough to angular particles) as well as for spherical particles with yield.
866 Observed values for β for terrestrial sands vary from 1/3 to 1/6 (e.g., Zimmer et al., 2007).

867 Calculating the increases of confining stress with depth corresponding to the three density curves
868 presented in Figure 3 leads to three corresponding velocity-depth profiles (Figure 10). However,
869 differences between the three profiles are barely distinguishable, which is to be expected given the
870 reported limited influence of density on the velocity increase with depth.

871 == Figure 10 about here ==

872 Equation 20 and the velocity measurement on the Mojave simulant have already demonstrated an
873 important application in modeling the different seismic noise sources that may affect the InSight
874 seismometers at various frequencies (Mimoun et al., 2017), although strictly speaking this model is only
875 sensitive to the shear modulus and Poisson's ratio as the model is mostly integrating noise sources from
876 static loading. For example, atmospheric pressure fluctuations on Mars induce an elastic response in
877 the ground creating ground tilt, detectable as a gravity signal on the InSight seismometer SEIS. The
878 amplitude of this pressure noise depends on the shear modulus and Poisson's ratio of the ground that
879 are related, and may be derived from the seismic velocities and an assumed bulk density (Murdoch et al,
880 2017a). A further example is dynamic pressure due to wind that results in stresses on the InSight lander
881 body and leading to ground deformation at the lander feet (Murdoch et al., 2017b). To calculate the
882 resulting ground deformation at the seismometer's ground position for a given wind dynamic pressure
883 and direction, local elastic properties beneath each foot of the lander are required. Seismic velocities
884 may be obtained from equation 20 by taking into account the pressure exerted by the lander mass
885 under Martian gravity and the elastic properties (shear modulus and Poisson's ratio) can then be
886 derived. The noise maps produced by Murdoch et al. (2017b), based on these calculations, will assist in
887 deploying SEIS at a site with little noise due to wind-induced ground deformation generated by the
888 lander.

889 For v_s , no relation corresponding to Equation (20) was derived by Delage et al. (2017). However, it
890 was found that the ratio between v_p and v_s remained rather constant for different confining stresses and
891 for the different simulants tested. Thus, the values of v_s shown in Figure 10 are derived from v_p using
892 the measured ratio of 1.669. The Poisson's ratio ν calculated via Equation (19) accordingly is 0.22
893 (Delage et al., 2017).

894 The velocity profiles in Figure 10 assume that the regolith is composed purely of sandy material.
895 Rock abundance at the landing site is low (see section 1.2), though, and a fraction of 5% or 10% rocks
896 would increase velocities v_p and v_s by less than 0.5% and less than 1.25%, respectively, for all three
897 models. This estimate is based on using the Reuss average, as in Delage et al. (2017), and assuming rock
898 properties of $v_p = 3000$ m/s, $v_s = 1700$ m/s and $\rho = 2760$ kg/m³ derived from terrestrial data obtained for
899 fractured basalt (Planke et al., 1999; Vinciguerra et al., 2005; Stanchits et al., 2006; Fortin et al., 2011) as
900 well as a negligible influence of compression on the rocks within the upper 5 m of the regolith. An
901 example of extending the velocity model to greater depths to include the coarse ejecta layer and the
902 transition from fractured to pristine basalt can be found in Knapmeyer-Endrun et al. (2017).

903 Terrestrial lab measurements on unconsolidated dry quartz sand result in P-wave velocities around
904 250 m/s and S-wave velocities around 150 m/s for confining stresses below 50 kPa (e.g., Velea et al.,
905 2000; Zimmer et al., 2002; Prasad et al., 2004). A terrestrial field experiment on soil with a low water
906 content yielded P-wave velocities as low as 150 m/s and S-wave velocities as low as 100 m/s directly at
907 the surface (Uyanik, 2010), whereas field measurements on beach sand showed P-wave velocities as low
908 as 40 m/s and an average of 160 m/s above the water table at 1.4 m depth (Bachrach et al., 1998). A
909 summary of terrestrial field results from exploration studies also finds P-wave velocities around 200 m/s
910 in shallow soils (Ohsaki and Iwasaki, 1973). Thus, the regolith velocity models are within the range
911 observed for terrestrial unconsolidated sands and soils.

912 The measured Poisson's ratio of 0.22 is low compared to values typically assumed for terrestrial
913 sediments. It is close to laboratory data for dry quartz sands: saturated sands show much larger
914 Poisson's ratios, in excess of 0.4, and corresponding v_p/v_s ratios up to and larger than 5 (Ohsaki and
915 Iwasaki, 1973; Prasad et al., 2004). The field experiment on beach sand also yielded a low Poisson's
916 ratio of 0.15 independent of depth (Bachrach et al., 2000). The field measurements by Uyanik (2010)
917 resulted in a v_p/v_s ratio of 1.5, corresponding to a Poisson's ratio of 0.1, for the upper tens of cms of dry
918 unconsolidated top-soil, indicating a porous and air-filled environment. These observations
919 demonstrate the strong influence of water content on Poisson's ratio in unconsolidated sands and soils.
920 As no free near-surface water is expected in the regolith at the landing site, but the layer is expected to
921 be porous and to exchange gases with the atmosphere, the low Poisson's ratio and v_p/v_s ratio
922 corresponding to values obtained from the laboratory experiments are plausible first estimates for the
923 InSight landing site.

924 For the Moon, seismic velocities at the surface initially derived from the touchdown of the Surveyor
925 spacecraft yielded very low values of 45 m/s for v_P and 23 m/s for v_S , corresponding to a Poisson's ratio
926 ν of 0.32 (Sutton and Duennebieer, 1970). Active seismic experiments of Apollo 14, 16 and 17 found
927 somewhat higher P-wave velocities of the lunar regolith of 100 to 114 m/s in the upper 4 to 12.2 m, with
928 higher velocities in the range of 250 to 330 m/s at greater depth (Kovach and Watkins, 1972; Watkins
929 and Kovach, 1972, 1973; Cooper et al., 1974). The v_P values for the uppermost regolith layer agree well
930 with estimates based on the recordings of the lunar module liftoff with the passive seismic experiments
931 at Apollo 12, 14 and 15, which are in the range of 99 to 103 m/s (Nakamura et al., 1975). Laboratory
932 measurements on lunar soils returned to Earth gave similarly low values for P-wave velocities of 125 m/s
933 at 4 kPa (Johnson et al., 1982). Gangi and Yen (1979) interpreted the data from the Apollo 14 and 16
934 active seismic experiments in terms of a power-law increase of P-wave velocity with depth in the
935 regolith layer, with an exponent of 1/6 as predicted by contact theory and a velocity of 110 m/s at the
936 surface, which was, however, contested by Watkins and Kovach (1973), claiming that this velocity law
937 does not provide a good fit to the layered Apollo models.

938 Shear wave arrivals were only tentatively identified in the active recordings of Apollo 14, resulting in
939 an S-wave velocity estimates of 62 m/s and a Poisson's ratio ν of 0.23 for the lunar regolith (Kovach
940 and Watkins, 1973), quite similar to the proposed model for the InSight landing site. Additional
941 information has been derived from the passive lunar experiments, e.g., horizontal-to-vertical spectral
942 ratios (H/V) of artificial and natural impacts as well as deeper events (Mark and Sutton, 1975; Horvath et
943 al., 1980). Lunar S-wave velocities were in the range of 32 to 40 m/s at the surface, with values greater
944 than 100 m/s found only below 10 m depth. Resulting Poisson's ratios are 0.41 to 0.43 at the surface,
945 decreasing to 0.33 below. Analysis of Rayleigh waves extracted from ambient noise correlations at the
946 Apollo 17 geophone array yielded S-wave velocity values of 50 m/s for the uppermost 2 m, increasing to
947 70 m/s at 4 m depth, and a Poisson's ratio around 0.33 (Larose et al., 2005; Sens-Schönfelder and
948 Larose, 2010). A recent re-analysis of horizontal to vertical spectral ratio (H/V) curves in combination
949 with Rayleigh wave dispersion from the active experiments at Apollo 14 and 16 yielded S-velocities of 50
950 to 60 m/s for the upper 12 to 15 m (Dal Moro, 2015), and Poisson's ratios around 0.33. In contrast, re-
951 analysis of Apollo 17 active seismic data using wavefield gradient analysis resulted in S-wave velocities of
952 40 m/s for the upper 4 m, underlain by 110 m/s. A Poisson's ratio ν of around 0.41 was indicated in the
953 shallowest layer (Sollberger et al., 2016).

954 While the low velocities of the lunar regolith are surely due to a high porosity (Tittmann et al., 1972),
 955 which also has a profound effect on velocities in terrestrial field experiments (Watkins et al., 1972), the
 956 vacuum does not play a major role. In experiments using a granular material consisting of glass beads,
 957 Griffiths et al. (2010) observed no difference between seismic velocities measured in vacuum and in
 958 ambient air, and only a relatively small decrease of a few percent for vacuum compared with 0.6%
 959 interstitial water, even at low confining pressure. In fact, the P-wave velocities for the shallowest layer
 960 measured during the Apollo program are in good agreement with terrestrial field measurements on
 961 sand and the predictions for the InSight landing site. The velocity law derived by Gangi and Yen (1979)
 962 predicts a much smaller increase of velocity with depth compared to the InSight landing site model
 963 (Figure 10). To a large extent, the resulting lower velocities at depth can be explained by reduced
 964 compaction under the diminished gravity of the Moon, although variations in grain size with depth
 965 might also affect the profile (Pilbeam and Vaišnys, 1973). Most measured lunar S-velocities are
 966 somewhat lower, and the Poisson's ratio accordingly higher, than predicted for the InSight landing site
 967 and found in dry terrestrial samples. However, the spread in v_s estimates, and correspondingly
 968 Poisson's ratio, for the lunar regolith is significantly larger than for v_p , which may explain part of the
 969 discrepancy.

970 4.2. Elastic Modulus

971 Based on Hooke's law, the elastic or Young's modulus E describes the ratio between uniaxial tensile
 972 stress σ and the proportional deformation, or extensional strain, ε , and thus the stiffness of a material:

$$\sigma = E\varepsilon \quad (21)$$

973 It can be expressed in terms of the shear wave velocity v_s , Poisson's ratio ν and density ρ as

$$E = 2v_s^2\rho(1 + \nu) \quad (22)$$

974 Depth profiles of Young's modulus for the three different models of regolith compaction are given in
 975 Figure 3. The values are lower than those obtained for some field tests on terrestrial soil, that found E
 976 increasing from 30 to 90 MPa in the upper 0.6 m (Uyanik, 2010), and on sand, that deduced E between
 977 20 to 70 MPa in the uppermost meter (Jaksa et al., 2004). In their overview, Bowles (1966) quote values
 978 between 5 and 25 MPa for E in silty to loose sand and a range of 50 to 81 MPa for dense sands, though,
 979 in good agreement with values calculated here. Teanby et al. (2016) also obtained low values for the
 980 effective E in the range of 1.1 to 4.4 MPa when applying elastic theory at two sites located on very loose

981 basaltic sands in Iceland. These values are likely appropriate only for the uppermost few centimeters of
982 the subsurface, whereas the profiles in Figure 11 show slightly larger values around 7.5 MPa.

983 == Figure 11 about here ==

984 In situ measurements of Young's modulus for the Moon were not reported but Alshibli and Hasan
985 (2009) determined E by laboratory experiments for the JSC-1A lunar regolith simulant, which is mined
986 from a volcanic ash deposit in a commercial quarry. They measured values in the ranges of 11.1 to 15.5
987 MPa and 10.3 to 27.6 MPa for loose and dense packing, respectively, at pressures corresponding to 2
988 and 4 m depth on Mars (10 and 20 kPa). These values are considerable lower than the values for E
989 calculated here, but JCS-1A has a large proportion of small grains, with more than 55% of grains smaller
990 than 100 μm . Thus, JSC-1A is not a good analogue of the regolith at the InSight landing site.

991

992 **4.3. Attenuation Factor**

993 Seismic attenuation is the dissipation of energy through internal friction and other non-elastic
994 processes and affects the amplitude of seismic signals propagating through natural materials.
995 Attenuation is quantified by the dimensionless seismic quality factor Q , defined via the decrease of
996 amplitude A at frequency f after travelling a distance x through a medium with seismic velocity v

$$A(x) = A_0 e^{-\left(\frac{f\pi}{Qv}\right)x} \quad (23)$$

997 (Lay and Wallace, 1995). Note that this equation defines attenuation caused by intrinsic anelasticity and
998 does not include apparent attenuation due to scattering, i.e., the redistribution of energy to the coda of
999 a seismic phase due to small-scale heterogeneity along the wave path. For the Moon, attenuation due
1000 to intrinsic anelasticity is much lower than on Earth, while scattering in the lunar crust is much larger,
1001 which, in combination, result in the characteristic signal shapes of lunar seismograms (e.g., Dainty and
1002 Toksöz, 1981). The envelope of these seismograms can be fairly well modeled by diffusion theory (see
1003 Lognonné et al., 2009; Gillet et al., 2017; for recent applications). No laboratory measurements of Q are
1004 available for Martian regolith analogues. Thus, the discussion is focused on available theories and on
1005 data from the Moon and Earth, which are clearly different, and what can be deduced from these for
1006 Mars.

1007 S-wave quality factors Q_S , obtained by borehole measurements in terrestrial sediments and soils, lie
1008 between 3 and 35 (e.g., Gibbs et al., 1994; Assimaki et al., 2008; Parolai et al., 2010; Fukushima et al.,
1009 2016). From surface measurements on Quaternary sediments, Malagnini (1996) determined a
1010 frequency dependence in Q for both P- and S-waves, with $Q_P = Q_S = 9$ at 10 Hz, compared to a value of 2
1011 previously found at 1 Hz (Malagnini et al., 1995). Frequency dependence in Q at frequencies of a few Hz
1012 is generally attributed to the influence of scattering (e.g., Kinoshita, 2008), which we do not consider
1013 further here. Jongmans (1990) found similarly low values, on the order of 5, for Q_P in field
1014 measurements on unsaturated sand. Laboratory measurements on dry quartz sands showed Q_S in the
1015 range of 15 to 50 at lowest confining pressures below 0.3 MPa and Q_P around 10 to 15 (Prasad and
1016 Meissner, 1992).

1017 In contrast to terrestrial data, Apollo experiments determined unusually high Q values in the lunar
1018 interior, ranging from 3000 to 3600 in the upper crust (Latham et al., 1970a, b) to 4000 to 4800 in the
1019 upper mantle for both P- and S-waves (Nakamura et al., 1976; Nakamura and Koyama, 1982). These
1020 high Q values also extended up to the near-surface material, including the lunar regolith and the
1021 somewhat faster layer below, for which Nakamura (1976) determined 2000 as a lower limit for Q from
1022 interpretation of rover signals. Analysis of the Apollo 14 seismic experiment data gave an estimate of
1023 50-100 for Q of the near-surface lunar material (Kovach and Watkins, 1972). Recently, Dal Moro (2015)
1024 found that high Q_S values of at least 100 in the uppermost regolith and 300 below the slowest layer to a
1025 few 100 m depth in the shallow crust are essential in obtaining a good fit to measured H/V curve
1026 amplitudes. As these data cannot differentiate further between Q_S values of either a few hundred or
1027 significantly larger (≥ 1000), they are not in conflict with previous higher estimates which averaged over
1028 larger depth ranges.

1029 As demonstrated in laboratory experiments, high Q values are caused by extremely low water
1030 content in the rocks from which even thin layers of adsorbed water have been removed by strong
1031 outgassing under vacuum conditions (Tittmann, 1977; Schreiber, 1977; Tittmann et al., 1979). As
1032 discussed by Tittmann et al. (1972), laboratory measurement of Q factors on returned lunar samples
1033 failed to reproduce the high values measured in situ on the Moon when exposing the samples to
1034 laboratory air during the measurements, and values around 50 to 100 were obtained. Only by
1035 outgassing the samples under high vacuum, could Q values of 3000 to 4500 be achieved, in agreement
1036 with the in situ estimates for lunar rocks. However, Q returned to the low original values after a few
1037 minutes re-exposure to laboratory air (Tittmann et al., 1979). However, all of these measurements

1038 pertain to lunar rocks, not fines. A similar observation was reported by Pandit and Tozer (1970) for
1039 porous terrestrial rocks, with an increase in Q by a factor of 5 between terrestrial atmospheric pressure
1040 and 1.5 Pa. Tittmann et al. (1980), working with porous sandstone, showed that the first monolayer of
1041 adsorbed water has the strongest effect and decreases Q by a factor of about 5 compared to the
1042 vacuum-dry case. In the Martian crust an evacuation of trapped fluids comparable to the lunar situation
1043 is prevented by atmospheric pressure, as it requires successive heating cycles at pressures below 1.5 Pa
1044 (Lognonné and Mosser, 1993). Accordingly, Q is predicted to be larger by at most a factor of two
1045 compared to Earth for Martian crustal rocks.

1046 A laboratory experiment on fines was conducted by Jones (1972). Jones used powdered basalt with a
1047 mean particle diameter of 5 μm and a mean density of 1340 kg/m^3 , significantly finer than the sand at
1048 the InSight landing site, but with a similar surface density to that estimated here. At 10 Hz Jones found a
1049 clear increase in Q with decreasing pressure, from values of Q_P around 50 at ambient conditions to 100
1050 at Mars surface atmospheric pressure, to 120 at about 5 Pa. Jones inferred that remnants of lubricating
1051 water films are still present at these pressures as compared to measurements made in a vacuum. For
1052 glass beads, 400-800 μm in diameter, Griffiths et al. (2010) reported differences in Q by a factor of 4.5
1053 between 200 in ambient air with about 25% humidity, and 900 in a vacuum. Brunet et al. (2008)
1054 obtained a Q of 295 for a similar granular material of glass beads, 600-800 μm in diameter, dried in a
1055 furnace, and measured under ambient conditions. According to contact theory for spherical particles,
1056 certain variables, including Q , are proportional to particle radius (Brunet et al., 2008), which could
1057 explain the different values obtained for Q in the different experiments.

1058 Laboratory measurements on dry quartz sand yield Q_P/Q_S ratios ranging from 0.2 to 1.8 (Prasad and
1059 Meissner, 1992; Prasad et al., 2004). Studies on porous sandstones yield equal values for Q_P and Q_S at
1060 low confining pressures when performing measurements under ambient laboratory conditions and after
1061 drying the samples in a laboratory oven (Toksöz and Johnson, 1979). Based on the limited information
1062 available, we assume that Q_P and Q_S are approximately equal at the InSight landing site.

1063 One of the main factors controlling Q is the regolith water content. Laboratory measurements have
1064 shown that a single monolayer of adsorbed water can drastically reduce the high Q values observed in
1065 outgassed lunar or terrestrial samples (Tittmann et al., 1979, 1980). Pandit and Tozer (1970) reported
1066 that the large change in Q they observed was connected to a change in water content of less than 0.05
1067 wt.%. Any liquid or frozen surface water would not be in equilibrium in the equatorial regions of Mars

1068 targeted by the InSight lander and would quickly sublimate (Golombek et al., 2017). However, water
1069 within the regolith could still be present in the form of a few monolayers of adsorbed water (Möhlmann,
1070 2008), which would maintain liquid-like properties down to temperatures of -70°C (Lorek and Wagner,
1071 2013). This adsorbed water is supposed to reside mainly below depths of a few tens of cm, outside the
1072 range of the Martian diurnal and seasonal thermal cycles (Möhlmann, 2004). Such a two-layered
1073 regolith structure would be consistent with a model for regolith water content derived from neutron
1074 spectroscopy data (Feldman et al., 2004), which assumes a relatively desiccated near surface layer with
1075 2 wt.% water and a more water-rich layer below, with at least 6 wt.% water. Furthermore, given that
1076 the Martian regolith is expected to be in exchange with the atmosphere (see Section 6 below), it seems
1077 reasonable to assume that monolayers of water could be present, but the amount of water in the
1078 regolith depends on latitude and season (Martinez et al., 2017). This would also be consistent with
1079 degassing experiments performed by the SAM (Sample Analysis on Mars) instrument suite on the
1080 Curiosity rover at Gale crater (Leshin et al., 2013), which found loosely bound water degassing from the
1081 samples starting at around 100°C .

1082 Therefore, we provide models for Q values for the Martian regolith that are based on Mindlin
1083 theory (Figure 12), as used by Brunet et al. (2008) to interpret their data from measurements with dry
1084 beads. The resulting values are consistent with results obtained in lab experiments on basalt fines and
1085 granular materials in dry, but non-vacuum conditions, taking into account estimates for regolith particle
1086 size. The theory predicts a dependence of Q on pressure with an exponent of $2/3$, which is within the
1087 observed range of 0.5 to 0.9 for spherical grains (Pilbeam and Vaišnys, 1973). Observations for angular
1088 grains found a smaller pressure dependence with an exponent of 0.3 to 0.4 (Pilbeam and Vaišnys, 1973).
1089 The increase of Q with depth could thus be lower if the particle grains at the landing site are less than
1090 perfectly spherical. In addition, Q also depends on particle size. We used a particle radius of $100\ \mu\text{m}$, in
1091 the center of the range for fine sand when calculating the curves in Figure 12. However, a non-uniform
1092 particle size will result in deviation in the predicted values for Q . Specifically, if particle size increases in
1093 the upper 5 m of the regolith, the increase in Q with depth will be larger. Finally, Mindlin theory also
1094 predicts an inverse dependence of Q on displacement amplitude which was not observed in some low
1095 pressure experiments (Pilbeam and Vaišnys, 1973). Here we consider amplitudes related to the low
1096 end-member strain analyzed by Brunet et al. (2008), on the order of 5×10^{-6} , to avoid decreasing Q .

1097 == Figure 12 about here ==

1098 The Q values estimated here are lower than some of the estimates for the lunar regolith, but
1099 distinctly higher than terrestrial values. However, it is worth repeating that if no adsorbed water is
1100 present in the Martian regolith, Both Q_P and Q_S could be larger than the values given here by up to an
1101 order of magnitude.

1102 Surface waves have their amplitude maximum at one-third of their wavelengths. Thus, short period
1103 surface waves with a period of 7 Hz, such as those observed in autocorrelations of Apollo 17 geophone
1104 data from the Moon, and a group velocity of about 100 m/s based on the estimates in Section 5.1, are
1105 strongly influenced by the regolith layer. The range of Q deduced here would indicate approximately 5
1106 to 6 s of propagation time for one Q cycle, or 500 to 600 m of propagation distance for these waves.
1107 Amplitude could be reduced by a factor of two after 500 to 600 m of propagation, limiting the
1108 observational range of the waves.

1109

1110

1111 **5. Mass Diffusivity**

1112 The section concerns the mass diffusivity, or coefficient of mass diffusion, of the Mars atmosphere
1113 with respect to the porous medium of the regolith at the InSight landing site. This parameter is
1114 important because the atmosphere flows in and out of the regolith in response to changes in
1115 atmospheric pressure, and has the potential to convectively transfer heat in and out of the regolith.
1116 Convective heat transport associated with atmospheric pressure changes could be indicated by
1117 transients in the HP³ temperature data and/or variations in calculated heat flow with depth. Mass
1118 diffusivity is somewhat analogous to thermal diffusivity where thermal diffusivity can be used to
1119 describe heat diffusion in the subsurface (see subsection 4.6 Thermal Diffusivity above). In a simplified
1120 form, effective mass diffusivity, D_{eff} , may be defined by the following equation (*cf.*, Scanlon et al., 2002,
1121 equation 8.31):

$$\frac{\partial M}{\partial t} = D_{eff} \frac{\partial^2 M}{\partial z^2} \quad (24)$$

1122 where M is mass of the diffusing gas, t is time, and z is depth. Unlike heat flow, however, in porous
1123 media the gas molecules flow through the pores rather than through the minerals grains (heat may also

1124 be transferred through pores by radiation). Gas molecules have random motion, influenced by pressure
1125 gradients, and their interactions with the minerals depend on the molecular gas mean free path, λ ,
1126 relative to the average pore radius, r .

1127 Mass diffusivity has been measured in terrestrial regoliths (soils and subsoils) under the same
1128 conditions of atmospheric pressure change as we are interested in Mars. Cyclic changes in atmospheric
1129 pressure that propagate into the subsurface are commonly known as *barometric pumping* or
1130 *atmospheric breathing*. On Earth they are of interest in studies of gas exchange associated with plant
1131 growth in the vadose zone and in studies of vertical transport of contaminated gases in the porous
1132 subsurface (*e.g.*, Nilson et al., 1991; Massmann and Farrier, 1992; Rossabi and Falta, 2002; Massmann,
1133 2006; Rossabi, 2006). These studies are applicable to barometric pumping on Mars at the macro scale,
1134 *i.e.*, in the pumping theory, but miss an important difference in the pressure diffusivity at the molecular
1135 scale between Earth and Mars. As a consequence of Mars' low atmospheric pressure, molecules in the
1136 regolith of Mars have a much higher mean free path than molecules in the terrestrial regolith. They
1137 interact more with the pore walls than with their neighboring gas molecules, whereas terrestrial gas
1138 molecules generally interact more with each other except in very fine-grained materials, such as shales.
1139 Terrestrial gas molecules in porous media interact with the pore walls when the pores are very small.
1140 Pore-wall interactions are important in terms of the permeability and pressure diffusivity of the Mars
1141 regolith, and are discussed below. There is one set of experimental measurements of pressure
1142 diffusivity under Mars surface atmospheric conditions (Fanale et al., 1982a): these results are discussed
1143 and compared with theoretical calculations after presentation of molecular gas interactions in porous
1144 media.

1145 **5.1. Gas Interactions in Porous Media**

1146 At low mass concentrations and in small pore passages, diffusion of gas molecules in porous media
1147 involves collisions between the gas molecules and the porous media in addition to molecular
1148 interactions among the gas molecules. Mass diffusivity and permeability are both parameters that
1149 relate to the flow of fluids through porous media, but they are not simply related because mass
1150 diffusivity includes the effects of compressibility, especially when the fluid is a gas (*e.g.*, Liang et al.,
1151 2001). However, some of the interactions among gas molecules with pore walls that apply to mass
1152 diffusivity were first studied and observed in permeability. One of the interactions of gas molecules
1153 with pore walls is slip of gas molecules near a solid wall. Klinkenberg (1941) first addressed how this

1154 interaction can affect the measured permeability of a gas, and he proposed a linear permeability
1155 correction. Four modes of diffusion have been described which are usually distinguished by the
1156 Knudsen number, K_n (e.g., Ziarani and Aguilera, 2012):

$$K_n = \frac{\lambda}{\delta} \quad (25)$$

1157 where λ is mean free path of the gas molecules and δ is a characteristic length, such as the pore
1158 diameter. Three of the four modes of diffusion are illustrated in Figure 13 and the four modes and their
1159 relations to the Knudsen number are described in Table 3.

1160 == Figure 13 about here ==

1161 == Table 3 about here ==

1162 For small Knudsen numbers that are applicable to most terrestrial gas flows in natural porous media,
1163 pressure diffusivity coefficients representative of Darcy flow are appropriate. However, as the Knudsen
1164 number increases to where slip flow on pore boundaries dominates, a new diffusion coefficient, the
1165 Knudsen diffusivity, is more accurate (see Table 3). The Knudsen diffusion coefficient, D_k , is given by
1166 (e.g., Huizenga and Smith, 1986; Roy et al., 2003; Javadpour et al., 2007):

$$D_k = \frac{\delta_p}{3} \sqrt{\frac{8RT}{\pi M}} \quad (26)$$

1167 where δ_p is the pore diameter R is the universal gas constant, T is absolute temperature, and M is the
1168 gas molar mass. Under conditions of Knudsen diffusion (Table 3, $K_n > 10$), D_k is the appropriate
1169 diffusion coefficient to use in Equation 24 in place of D_{eff} .

1170

1171

1172 **5.2. Estimating Pore Sizes**

1173 Many variables contribute to the pore radii in sediments and porous rocks, including grain size,
1174 degree of sorting, compaction, cementation, moisture content, diagenesis, and growth of secondary
1175 minerals. There is evidence of wind and water processes on the surface of Mars, both of which would

1176 tend to sort and round grains in the regolith. Impact processes produce angular fragments and poorly
1177 sorted materials. The landing ellipse for the InSight landing site was chosen to be on smooth, flat terrain
1178 that generally has a very low rock abundance and as few impact craters visible in high-resolution orbital
1179 images as possible (Golombek et al., 2017). Selection criteria for the landing site in the northern
1180 lowlands and with a paucity of impact craters should make impact fragmentation subordinate to
1181 abrasion as a mechanical weathering process at the landing site. The particles in the landing site
1182 regolith may therefore be expected to be well-sorted, rounded grains, as described in Section 2 above.

1183 Although relations have been proposed, no universal simple relation exists in sediments between
1184 grain size and pore radii from which the pore radii may be estimated. Kaviany (1994) proposed a
1185 relation among average pore size, particle diameter and porosity for spherical particles in random
1186 packing. If a fractional porosity of 0.399 is assumed, representative of random packing of uniform
1187 spheres, this relation gives a ratio of average pore size to grain size, δ_p/d_g , of 0.072, where δ_p is pore size
1188 and d_g is the grain diameter. Minimum pore throat diameters were calculated geometrically assuming
1189 the most inefficient regular packing of uniform spheres (Cubic packing, 0.476 porosity), and the most
1190 efficient regular packing of uniform spheres (Triclinic, or hexagonal close packing, 0.260 porosity). For
1191 cubic packing the minimum throat diameter is given by $0.207d_p$ ($\delta_p/d_g = 0.21$); for triclinic packing the
1192 minimum throat diameter is given by $0.0774d_p$ ($\delta_p/d_g = 0.077$). Assuming a porosity representative of
1193 random packing, the ratios of pore diameter or pore throat diameter to grain size (δ_p/d_g) calculated from
1194 the from the Kaviany (1994) equation are very similar to those calculated geometrically for triclinic
1195 (close-hexagonal) packing, 0.072 versus 0.077, respectively. Cubic packing is improbable in sorted
1196 spherical grains as they are unlikely to be balanced in vertical columns.

1197 One further complication in determining pore size from grain size is that the methods discussed
1198 above all assume uniform spherical grains, a condition that may not exist in the Mars regolith.
1199 Variations in grain size and deviations from spherical shape are both likely to reduce pore size as smaller
1200 grains would fill larger pore spaces and flattening of the grains would result in compaction: reduced
1201 pore size would reduce pressure diffusivity. However, at the InSight landing site the surface regolith
1202 sediment is likely to be well-sorted and rounded from eolian processes. Using the estimated range of
1203 grain size of 0.125 to 0.25 mm (radii 0.0625 to 0.125 mm) from Subsection 2.1 Landing Site Overview
1204 above, and an average δ_p/d_g ratio of 0.075, a range of pore throat diameters of 9.4 to 18.8 μm was
1205 calculated. At the InSight landing site these pores would be subject to an atmospheric pressure range of
1206 6 to 8.5 hPa.

1207 **5.3. Gas Mean Free Path and Range at Landing Site**

1208 The mean free path of molecules in a gas is estimated by considering the volume of a cylinder that
1209 represents the gas molecules effective collision area, including the area of target molecules in this area,
1210 with respect to the distance travelled by the molecules and the number of molecules per unit volume
1211 (*e.g.*, Nave, 2016). The number of molecules per unit volume of gas may be approximated by assuming
1212 that the systems behaves as an ideal gas (Tan, 2014). The calculation must also recognize that both the
1213 colliding and the target molecules are moving (Nave, *op. cit.*). These assumptions yield the result that
1214 the molecular mean free path, λ , may be estimated by:

$$\lambda = \frac{RT}{\sqrt{2}\pi\gamma^2 N_a P} \quad (27)$$

1215 where R is the universal gas constant, T is absolute temperature, γ is the effective collisional diameter of
1216 the molecules, N_a is the Avogadro number, and P is pressure. The effective collisional diameter of CO₂ is
1217 330 pm (*e.g.*, Albrecht et al., 2003), and at a temperature of 180 K and pressures of 6 and 8.5 hPa,
1218 molecular mean free paths of 8.56 and 6.04 μm were calculated for CO₂. At a temperature of 270 K and
1219 pressures of 6 and 8.5 hPa, molecular mean free paths of 12.8 and 9.06 μm were calculated. This array
1220 of conditions and calculated molecular mean free paths should cover the range of likely diffusivity
1221 environments to be encountered at the InSight landing site.

1222

1223 **5.4. Calculated Range of Mass Diffusivity at Landing Site**

1224 Knudsen numbers were calculated using the molecular mean free paths calculated with equation 27
1225 for the range of pore diameters estimated above, and corresponding Knudsen diffusion coefficients
1226 were calculated using equation 27. These results indicate that gas flow in the shallow regolith at the
1227 InSight landing site will probably be in the Knudsen Transition Flow range with Knudsen diffusivities
1228 ranging from of 1 to 2 x 10⁻³ m²/s. To give a direct comparison of Knudsen diffusivity with grain size
1229 when in the pore and pressure range for which the Knudsen diffusivity equation is applicable, Knudsen
1230 diffusivity is plotted as a function of grain size in Figure 14 for the expected range of grain sizes for the
1231 near-surface regolith at the InSight landing site.

1232 == Figure 14 about here ==

5.5. Comparison with Experimental Data

Fanale et al. (1982a) built an experimental system to determine the mass diffusivity of a *Mars simulant soil* (45% smectite, 45% finely-ground basalt, and 10% iron oxide) with a density of 1300 kg/m^3 at temperatures of -40°C (233 K) and -70°C (203 K). Diffusivity was determined by measuring the rate of penetration of a CO_2 pressure wave with a starting pressure of $\sim 6 \text{ hPa}$ and a pressure step of $\sim 2 \text{ hPa}$. The experimentally estimated diffusivities were 2.5×10^{-6} and $1 \times 10^{-6} \text{ m}^2/\text{s}$ for temperatures of 233 and 203K, respectively. Fanale et al. (1982a) did not give an estimate of the average pore diameter of their Mars simulant soil, but presumably the pores were very small as 90% of the simulant was smectite and finely-ground basalt. Their determined diffusivity range is three orders of magnitude smaller than the diffusivities calculated above. The primary difference in the diffusivities determined experimentally and the diffusivities calculated here may be explained by the smaller pore sizes in the experimental regolith simulant.

An additional phenomenon, discussed by Fanale et al. (1982b), is the adsorption of CO_2 onto the grains of the regolith. The adsorption of gases, including CO_2 , onto the surface of clays had been previously reported (*e.g.*, Aylmore et al., 1970; Fanale and Cannon, 1979). The adsorption of molecules onto grain surface tends to decrease pore diameters but does not reduce slip flow as molecules can slip over molecules adsorbed onto grains. CO_2 molecules are less than $0.001 \mu\text{m}$ in their longest dimension which much smaller than the pore sizes discussed above (9.4 to $18.8 \mu\text{m}$). Thus, even if several layers of CO_2 molecules adhere to the pore walls the reduction in pore size would be small. The effect would be to increase the Knudsen number, but it would be unlikely to move out of the transition flow mode, with a small accompanying decrease in mass diffusivity. These effects are likely to be very minor: a $0.01\mu\text{m}$ (10-20 layers of CO_2 molecules) reduction in the minimum pore size ($9.4 \mu\text{m}$) would result in a 0.11% increase in the Knudsen number and a 0.11% reduction in the Knudsen diffusivity. Adsorption of CO_2 could also impact the mass diffusivity by acting as a temporary reservoir for CO_2 , storing CO_2 by adsorption during pressure increases and releasing the adsorbed CO_2 during pressure decreases. This effect could result in a hysteresis in atmospheric breathing that could be complicated by the temperature sensitivity of adsorption.

5.6. Final Observations

Mass diffusivity is an important parameter to the InSight mission because it constrains the flow of the Mars atmosphere into and out of the regolith at the landing site in response to changes in

1263 atmospheric pressure. This is a well-known phenomenon on Earth. Although the pumping process is
1264 similar on Mars to Earth, the molecular processes controlling mass diffusivity are different as a
1265 consequence of the low pressure of the Mars atmosphere: on Earth gas molecular collisions are
1266 dominantly with neighboring gas molecules; on Mars gas molecular interactions are dominantly with
1267 regolith grain surfaces. Using a calculated range of pore sizes based on the assumption of uniform-size,
1268 spherical grains at the landing site, a range of mass diffusivities of 1 to $2 \times 10^{-3} \text{ m}^2/\text{s}$ was calculated. This
1269 is probably a high estimate as grains of variable size and non-spherical grains would generally result in
1270 smaller pores than uniform-size spherical grains. The calculated diffusivity range based on simplified
1271 grain geometry is significantly higher than an experimentally determined range of mass diffusivities for
1272 the Mars regolith of 1×10^{-6} to $2.5 \times 10^{-6} \text{ m}^2/\text{s}$ (Fanale et al., 1982a). A probably explanation for the
1273 difference between the calculated and experimentally determined diffusivity ranges is that the regolith
1274 simulant used by Fanale et al. (1982a) in their diffusivity determination was very fine grained. The
1275 inclusion of 45% smectite, a clay, in their sample suggests that at least part of their sample had a grain
1276 size in the range of ~ 0.1 to $0.4 \text{ }\mu\text{m}$. Assuming the same pore size to grain size as used above, a range of
1277 Knudsen numbers equivalent to the curves in Figure 14 of 20 to 43 was calculated corresponding to
1278 Knudsen diffusivity range of 1.0 to $2.6 \times 10^{-5} \text{ m}^2/\text{s}$ for 203 K and 1.1 to $2.6 \times 10^{-5} \text{ m}^2/\text{s}$ for 233 K. These
1279 results are about an order of magnitude lower than the diffusivities estimated experimentally, the
1280 differences probably being caused by the assumption of uniform spherical grains in the pore size
1281 approximation for the calculations: clays have platy grains and the average pore sizes in the
1282 experimental mixture were likely to be smaller than assumed here resulting in a lower experimental
1283 diffusivity. However, a grain size range of 0.125 to 0.25 mm and the calculated effective mass diffusivity
1284 with this grain-size range is thought to be more representative of the InSight landing site.

1285 What are the implications of the calculated mass diffusivities for the penetration of periodic
1286 atmospheric pressure waves into the regolith at the landing site? If we make the assumption that the
1287 regolith is homogeneous and isotropic, a penetration skin depth δ can be calculated as $\delta = \sqrt{(2\pi D_{eff} / \omega)}$,
1288 where ω is the angular frequency of the period wave. The skin depth is the depth at which the
1289 maximum amplitude of the pressure change is $1/e$ ($\sim 37\%$) of the maximum surface pressure change.
1290 For a wave with a period of 1 sol (24 hours 40 min), $\delta = 9.4 \text{ m}$ for $D_{eff} = 1.0 \times 10^{-3} \text{ m}^2/\text{s}$, and $\delta = 13.2 \text{ m}$ for
1291 $D_{eff} = 2.0 \times 10^{-3} \text{ m}^2/\text{s}$. For a wave with a period of Mars year (687 days), $\delta = 244 \text{ m}$ for $D_{eff} = 1.0 \times 10^{-3}$
1292 m^2/s , and $\delta = 345 \text{ m}$ for $D_{eff} = 2.0 \times 10^{-3} \text{ m}^2/\text{s}$. These are large depths relative to the maximum
1293 penetration of the HP³ probe of 5 m. The time for a diffusive disturbance to travel a characteristic

1294 length L_c of 5 m is about 0.29 sol for a diffusivity of $1.0 \times 10^{-3} \text{ m}^2/\text{s}$, and about 0.14 sol for a diffusivity of
1295 $2.0 \times 10^{-3} \text{ m}^2/\text{s}$ (using the approximation $L_c^2 = D_{eff} t$, where t is time). However, the effect of flow of
1296 atmospheric gases in and out of the regolith in terms of heat transport and the HP³ heat-flow
1297 determination depends on the relative efficiencies of convective gas heat transport and conductive heat
1298 transport (possibly aided by intergranular radiative heat transport. This problem has been examined by
1299 Morgan et al. (2017). Their highest estimate of mass diffusivity was an order of magnitude lower than
1300 we have concluded here for the regolith at the InSight landing site, but they concluded that the
1301 diffusivity would need to be higher by a factor of about 100 for convection to be more efficient than
1302 conduction with reasonable estimates of the thermal conductivity of the regolith. This conclusion is
1303 based on several estimated parameters, but current information indicates that atmospheric gases will
1304 be forced into the regolith by changes in atmospheric pressure, but thermal convection by these
1305 movements will be insignificant.

1306

1307 **6. Summary and Conclusions**

1308 There were a number of primary engineering criteria for the InSight landing site which to some
1309 extent affected the physical properties of the landing site. These criteria included latitude (equatorial
1310 for solar power), low elevation (avoid cold temperatures), smooth plains with few rocks and craters
1311 (safe landing site), and fragmented regolith (to be penetrated by the self-hammering, heat-flow probe –
1312 HP³). These criteria resulted in the selection of a 130 x 27 km landing ellipse at 4.5°N, 135.9°E in
1313 western Elysium Planitia on Hesperian plains in the southernmost lowlands.

1314 Thermophysical properties used in the site-selection process indicated a regolith at this site similar to
1315 weakly-bonded terrestrial soils, capable of being penetrated by the HP³ probe. The properties indicated
1316 that the soil was cohesionless sand or low cohesion soil with a bulk density of $\sim 1,000$ to $1,600 \text{ kg/m}^3$ and
1317 grain sizes of ~ 0.15 - 0.25 mm (fine sand). A cover of surficial dust was indicated, less than 1-2 mm thick,
1318 and with low rock abundance. The upper 5 m of the regolith were predicted to be composed of nearly
1319 cohesionless, fine, well-sorted, rounded to sub-rounded, basaltic sand, which included few rocks.

1320 Based on studies of terrestrial soils and from heat-flow observations on the Moon, the regolith
1321 density is likely to significantly increase with depth as a result of compaction. The lunar heat-flow
1322 results required a rapid increase in thermal conductivity associated with compaction with depth.

1323 Compaction caused by gravity and impacts have resulted in models based on lunar compaction but the
1324 models are uncalibrated for Mars.

1325 Information covering cohesion of the Mars Regolith at the InSight landing site has been compiled
1326 from mechanical arms from Mars landers and the wheels of rovers. Cohesions range from cohesionless
1327 to weakly cohesive soils, less than 4 kPa, with blocky soils having higher cohesions of 3-11 kPa. The
1328 landing site will probably have a thin layer of cohesionless to weakly cohesive eolian deposits at the
1329 surface. These deposits may be blown away by the pulsed jets of the lander, below which the regolith
1330 will be weakly cohesive.

1331 Internal friction angle is sensitive to factors including material grain shape and bulk density. Many
1332 Mars regolith simulants have had angular grains that are probably not representative of the rounded to
1333 sub-rounded grains subject to wind erosion at the landing site. Extrapolation of experiments with
1334 rounded grains and a bulk density of 1,300 kg/m³ have provided a friction angle of 28° to 30° for the
1335 landing site. If the assumption is made that particle shape does not change with depth, internal friction
1336 angle may be predicted as a function of bulk density and depth.

1337 Grain size is an important factor in many physical properties and is primarily constrained to be in the
1338 range of 150-250 μm (fine sand) by the thermal inertia of the landing site. Theoretical studies and
1339 observations at the Phoenix landing site in the Martian Arctic indicate that there is a transition below
1340 600 μm from larger clasts to the dominant fine sand grain size. Finer material may be found in this
1341 surficial dust layer.

1342 At this stage, thermophysical properties have been assumed to change only with depth.
1343 Measurements of surface emissivity on Mars has been from satellite sensors and from a sensor on the
1344 Mars Science Laboratory rover. These data have allowed weighted average emissivities to be derived
1345 for the three wavelength bands corresponding to the HP³ radiometer filters at 235 K for four different
1346 types of soils measured *in situ* by the Mars Exploration Rovers' mini-thermal emission spectrometer
1347 instruments.

1348 Surface thermal inertia controls the rate of change in temperature of the upper 2-30 cm of the
1349 regolith, and is strongly related to the square root of thermal conductivity. The lowest thermal inertias
1350 in the landing region are typically observed where atmospheric dust and very fine sand are trapped; the
1351 highest thermal inertias are associated with coarse regolith on crater rims and ejecta blankets.

1352 Surface albedo from different areas of Mars has been measured at different resolutions from orbiting
1353 satellite systems. Landers with retropropulsive thrusters have changed the surface albedo by
1354 temporarily removing the surface dust layer at all landing sites where the thrusters have been used. A
1355 temporary albedo reduction of ~20-50% at the InSight landing site during landing is anticipated.

1356 Based on in situ determinations of the thermal conductivity of the lunar regolith during two of the
1357 Apollo missions, and a number of published experiments simulating lunar and Mars regolith conditions,
1358 the thermal conductivity of the shallow regolith at the landing site is anticipated to be of the order of
1359 0.01 W/(m K), about two orders of magnitude lower than the thermal conductivity of damp terrestrial
1360 soils. As bulk density changes with depth, thermal conductivity is anticipated to change with depth. In
1361 addition, although atmospheric pressure is much lower, the fractional changes in atmospheric pressure
1362 during the diurnal and annual cycles are much greater on Mars than on Earth. As heat transfer through
1363 the gas in pore spaces is significant on Mars, the bulk thermal conductivity is sensitive to changes in
1364 atmospheric pressure.

1365 Studies of the heat capacity (units J/K) or specific heat (units J/(kg K)) of lunar, geologic, and
1366 meteorite materials at low temperatures indicate that these parameters are strongly temperature
1367 dependent, increasing with increasing temperature. This temperature dependence is most significant in
1368 the near-surface regolith layer where there are large temperature perturbations associated with diurnal
1369 and annual temperature variations. Below a few tens of cm these perturbations decay and an average
1370 heat capacity/specific heat may be used.

1371 Thermal diffusivity is the parameter in thermal conduction associated with the propagation of
1372 temperature changes, such as transmission of the annual temperature variation into the regolith. As
1373 with other thermal parameters, it is probably most variable in the upper few tens of cm of the regolith
1374 at the landing site, and is fairly constant below this depth.

1375 Subsurface elastic properties are of particular importance to the data to be collected by the
1376 seismometer experiment (SEIS) when operating at its highest rate and for short period surface waves
1377 above 5 Hz. There are no remote sensing data or existing lander results from which these properties
1378 may be derived and thus at present they are estimated from laboratory measurements. Seismic body
1379 wave measurements indicate that seismic velocities are very slow within the regolith but a significant
1380 increase in velocities may be expected between the surface and 5 m depth. In contrast, experiments on
1381 Mars regolith simulants and similar materials indicate that Poisson's ratio will be relatively constant with

1382 depth in dry, shallow regolith, but lower than most estimates for the Moon or measured in water-
1383 saturated terrestrial soils. Young's Modulus increases rapidly with depth, similar to the body-wave
1384 velocities. Seismic attenuation (dissipation of seismic energy by non-elastic processes), as measured by
1385 the seismic quality factor, Q , is expected to be relatively high in the Mars regolith, but depends to a
1386 large extent on the presence of adsorbed water, a parameter for which there are no direct observations
1387 at the InSight landing site. Q was measured to be very high, both in the regolith and at depth, on the
1388 Moon relative to terrestrial values, reflecting the very dry state of the Moon. A very small amount of
1389 water, monolayers in thickness, on the grains in the Mars regolith could be sufficient to significantly
1390 reduce Q by an order of magnitude, however. If no water is present Q would be close to lunar values.

1391 Mass diffusivity of the landing site regolith is the parameter that relates the flow of the Mars
1392 atmosphere in and out of the regolith in response to changes in surface atmospheric pressure. Most
1393 landing site physical parameters change from Earth to the Mars regolith because of differences in water
1394 saturation, atmospheric pressure, compaction, composition, etc. Mass diffusivity changes from Earth to
1395 Mars, except in a few special terrestrial examples, in that the mode of gas transport is dominated by
1396 molecule-grain collisions in the landing site regolith and a mass diffusivity equation appropriate to this
1397 mode (Knudsen diffusivity) must be used. The results of one experiment to measure mass diffusivity
1398 have been published, but the grain size of the material used in this experiment was much smaller than
1399 is thought to apply to the landing site. However, when the grain size and shape are included in
1400 estimation of the pore size, the calculated Knudsen diffusivity is close to the experimental results. The
1401 effective mass diffusivity calculated for the landing site is three orders of magnitude larger than the
1402 experimental results, but consistent with different grain size and shape.

1403 Physical properties of the regolith at the InSight landing site presented here are all speculative.
1404 Some of the properties are based on circular reasoning because they are based on data that were used
1405 to select the landing site, such as surface thermophysical properties. However, even these properties
1406 are ultimately based on correlations of remote sensing properties (satellite or rover) with ground truth
1407 data. Many of the properties are based on extensive experimental data with carefully refined models
1408 for the Mars regolith. However, with the exception of a shallow trench dug by the Phoenix lander in the
1409 southern polar region, and extrapolations from limited cliff exposures, there are no direct stratigraphic
1410 data describing the Mars regolith. We will gain much of these data during the penetration of the HP³
1411 probe and from the data collected during the InSight mission.

1412

1413 **7. References**

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1839 **10. Appendix**

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== Appendix table about here ==