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- We model seismic signals generated by meteorite impacts on Mars.
- We predict the total seismic moment released by impacts is $1e13$ - $1e14$ Nm per year.
- Nominally, about one globally detectable impact event should occur every ten years.
- Meteorite impacts do not provide a dependable way of probing the deep interior.

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Seismic detection of meteorite impacts on Mars

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Abstract

Meteorite impacts provide a potentially important seismic source for probing Mars' interior. It has recently been shown that new craters can be detected from orbit using high resolution imaging, which means the location of any impact-related seismic event could be accurately determined - thus improving the constraints that could be placed on internal structure using a single seismic station. This is not true of other seismic sources on Mars such as sub-surface faulting, which require location using multiple seismic stations. This study aims to determine the seismic detectability of meteorite impacts and assess whether they are a viable means of probing deep internal structure. First, we derive a relation between crater diameter and equivalent seismic moment based on observational data compiled from impact tests, controlled explosions, and earthquake seismology. Second, this relation was combined with updated cratering rates based on newly observed craters to derive the impact induced seismicity on Mars, which we estimate to total 10^{13} – 10^{14} Nm per year. Finally, seismic waveform modelling was used to determine the detectability of these impacts based on reasonable assumptions about likely seismometer performance and background noise levels. For our

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nominal noise/instrument case we find that detectable impacts at teleseismic distances (source-receiver offsets greater than 60°) are very rare and occur approximately once every 10 years. For our most optimistic noise/instrument case, approximately one such event occurs each year. This suggests that using solely meteorite impacts is not a reliable way of probing the Martian interior, although local impacts are more frequently detectable and could provide important constraints on near surface seismic properties.

Keywords: Mars, Seismology, Internal Structure, Planets.

1. Introduction

The major geophysical tools for probing planetary interiors are seismology, geomagnetism, observed orbital changes of planetary satellites, and measurements of the gravitational field and moment of inertia. Internal models can also be predicted using geochemical modelling based on the composition of representative meteorites. On Earth, seismology provides the most detailed information on internal structure and has been used to determine core and mantle structure (e.g. Dziewonski and Anderson, 1981). Further properties can be determined by measurements of the global magnetic field, especially the state of the conducting inner and outer cores.

Mars has no internal magnetic field (Acuna et al., 1999; Connerney et al., 1999) or reliable seismic data so currently the only way to estimate internal structure is to develop models that are consistent with the measured moment of inertia of $I=0.365MR^2$ (Yoder et al., 2003; Sohl et al., 2005) and an assumed composition based on SNC meteorites, which are believed to be representative of the bulk Martian composition (Sohl and Spohn, 1997). Many

different interior models can fit these constraints and as a result, basic properties remain relatively unconstrained. The size of the core is uncertain and there are currently at least three suggested core models with various combinations of solid and liquid cores (e.g. Stevenson, 2001). The composition and velocity structure of the core and mantle are also poorly constrained.

The best way to improve our knowledge of Mars' interior is with a global seismic network. However, off-Earth seismology is extremely challenging. The seismometer must survive the rigours of launch, landing, and harsh environments with large temperature swings. Another challenge is remote deployment onto the planet surface in a way that provides good seismic coupling. The Apollo seismic experiment, deployed by astronauts, detected numerous seismic events on the Moon, and has revolutionised our understanding of the lunar interior. Internal models now exist for lunar density (Bills and Ferrari, 1977), velocity structure (Goins et al., 1981; Nakamura, 1983; Khan et al., 2000; Lognonne et al., 2003; Gagnepain-Beyneix et al., 2006; Khan et al., 2007), and seismic attenuation (Nakamura, 1976; Goins et al., 1981; Nakamura and Koyama, 1982).

On Mars, the Viking seismometer (Anderson et al., 1976) was badly affected by wind noise due to its position on the lander and poor coupling to the ground, but was able to place crude upper limits on Mars' current activity. Anderson et al. (1976) suggest that the Viking seismometer would require a magnitude 6.5-7.0 event to be globally detectable (equivalent seismic moment 10^{19} – 10^{20} Nm), whereas Goins and Lazarewicz (1979) suggest a magnitude as high as 9.0-10.0 would be required. However, subsequent research has predicted current activity should be well below these levels.

Possible sources of seismicity on Mars include: (1) Fracturing caused by thermal and lithostatic stresses predicted from visible faults in high resolution images and topographic data (Golombek et al., 1992; Knapmeyer et al., 2006). These authors predict total yearly moment releases of 10^{18} Nm and 3×10^{16} – 5×10^{18} Nm respectively, occurring mostly as small sub-magnitude 5 events; (2) Volcanic events, for which there is evidence of recent (< 50 Ma) activity from crater counting. Sub-surface events are most likely as no current surface activity is observable; and (3) Meteorite impacts, which based on extrapolation of Apollo lunar data should be detectable (Davis, 1993). On Earth, the majority of seismic events are due to plate tectonics - a process that appears to be long since dormant on Mars - but which may have operated in Mars' early history prior to 3.5 Ga (Sleep, 1994).

Although Mars is much less seismically active than the Earth, the background noise is also expected to be orders of magnitude lower, due to the thin atmosphere coupled with the absence of major terrestrial noise sources such as ocean waves, anthropogenic sources, and lack of vegetation for coupling wind noise into the subsurface. Therefore, a seismic investigation of the Martian interior should be possible with a low noise - well coupled instrument network.

Both ESA and NASA are considering deployment of seismic instrumentation on Mars in the near future. Mission scenarios have included both broad-band and short-period instruments (Lognonne et al., 1996, 2000; Pike et al., 2005). Although Mars is predicted to be seismically active, this has not yet been confirmed with seismic instruments. Therefore, the initial goal of such a mission would be to determine the level and global distribution of

67 current Martian seismic activity. This is a key measurement for determining
 68 both the level of further seismic exploration and the ability of seismology to
 69 address the major goal of constraining Mars' internal structure.

70 Through model refinement by comparison with the observed phases a
 71 single seismic station would provide some new constraints on Mars' interior.
 72 However, such analysis will require many assumptions and accurate location
 73 of the seismic source would require triangulation from at least three or four
 74 well spaced stations (Mocquet, 1999). This is especially true for stress related
 75 faulting and sub-surface volcanism, for which the source location would be
 76 unknown. However, Malin et al. (2006) recently showed that it is possible
 77 to detect new meteorite impact sites on Mars - with 20 new impact craters
 78 catalogued over a seven year period. Therefore, if a similar standard of
 79 impact monitoring could be achieved during a seismic experiment, event
 80 location would be accurately known (to within meters, depending on the
 81 instrument pixel size) and the constraints we could place on Mars' interior
 82 would be greatly improved.

83 Impacts of meteors and projectiles have been detected on the Moon by
 84 the Apollo seismometers and suggest that seismic detection would be pos-
 85 sible (Oberst and Nakamura, 1987; Latham et al., 1970a,b). Davis (1993)
 86 extrapolated the Apollo results to Mars and concluded that meteorites may
 87 indeed act as a very effective seismic source, with around 20 globally de-
 88 tectable impacts occurring per year. However, the Moon is thought to be
 89 seismically very different to Mars with low velocity zones, low seismic attenu-
 90 ation, and significant sub-surface scattering. This leads to non-conventional
 91 waveforms comprising very long wave trains with extended P and S-wave

92 coda (Lognonne, 2005). The Moon also has no atmosphere, and lies in a
 93 different gravitational environment to Mars so the meteorite flux is expected
 94 to be very different. Davis (1993) tried to take many of these factors into ac-
 95 count, but never-the-less, extrapolating the lunar results to Mars introduces
 96 a large degree of uncertainty.

97 In this paper we use new craters observed by Malin et al. (2006) to re-
 98 move as many assumptions as possible, and directly model wave propagation
 99 through Mars' interior using a representative seismic model. From this we es-
 100 timate the number of observable meteorite impacts per year given estimates
 101 of seismometer sensitivity and Martian background noise.

102 **2. Mars' Meteorite Induced Seismicity**

103 Comparison of recent high resolution imaging of Mars from Mars Global
 104 Surveyor (MGS) over a seven year period has resulted in the detection of 20
 105 new impact craters with diameters ranging from 2 to 148 m (Malin et al.,
 106 2006). If we can determine the relation between observed crater size and the
 107 impact equivalent seismic moment for Mars then these observations can be
 108 used to determine the frequency of impacts with a particular moment. This
 109 defines the meteorite induced seismicity and determines how many meteorite
 110 strikes could be detectable in a given observation period. This should provide
 111 a more reliable estimate of meteorite detectability than scaling the Apollo
 112 seismic data.

113 To determine the seismicity we must consider four things: (1) The current
 114 impact rate on Mars for a given crater size. (2) The relation between observed
 115 crater sizes and impact energy. (3) The efficiency of conversion of impact

116 energy into seismic wave energy. (4) The relationship between seismic energy
 117 and an equivalent seismic moment. Each of these factors is now considered
 118 in turn in order to derive an overall relation between crater diameter and
 119 seismic moment.

120 *2.1. Current Impact Rate on Mars*

121 Hartmann (2005) presents the latest chronology of Mars' cratering his-
 122 tory. Assuming this chronology and that the cratering rate has been ap-
 123 proximately constant for the last 3 Gyrs gives the 1-year isochron shown
 124 Figure 1. However, considerable uncertainty exists in the small crater diam-
 125 eter (<1 km) regime and Hartmann (2005) quotes a factor of 3 uncertainty.

126 The 20 new craters observed by Malin et al. (2006) can be used to reduce
 127 this uncertainty and validate/adjust the chronology of Hartmann (2005) in
 128 this uncertain region. However, as discussed by Hartmann (2007) it should
 129 be noted that: (1) the frequency of occurrence of the smallest crater sizes
 130 observed by Malin et al. (2006) should be regarded as a lower limit, as the
 131 detection limit of their image analysis was around 10 m diameter; and (2)
 132 the largest new crater (148 m) was much larger than any of the other craters
 133 observed and is probably not representative of the time averaged meteorite
 134 flux.

135 Figure 1 compares the Malin et al. (2006) MGS results to the isochron
 136 of Hartmann (2005). The agreement is impressive considering the gross ex-
 137 trapolation required to get from the 3 Gyr to 1 year isochron. However, the
 138 MGS results suggest less cratering than predicted, and we find that the best
 139 fit to the observations is obtained by scaling the Hartmann (2005) isochrons
 140 by $1/3$ (if craters <10 m and the 148 m crater are ignored). This is also

141 consistent with Quantin et al. (2007), who suggest a factor of 3 reduction in
 142 impact rates over the last 3 Gyrs.

143 The Hartmann (2005) isochrons include a reduction in smaller impact
 144 events caused by atmospheric ablation and fragmentation effects discussed
 145 by Popova et al. (2003), who found that below 4 m the possibility of crater
 146 formation is heavily dependent on meteorite composition. Hence, we do not
 147 consider craters under 4 m diameter, as these will be difficult to detect from
 148 orbit and have a large uncertainty in flux.

149 *2.2. Crater Diameter-Impactor Energy Dependence*

150 Cratering mechanisms are reviewed in detail by Melosh (1980, 1989), and
 151 more recently by Richardson et al. (2005) and Holsapple and Housen (2007)
 152 motivated by the Deep Impact mission. Here, we are concerned with the
 153 relation between impactor kinetic energy E and the resulting crater diameter
 154 D . It is found that the relationship between these quantities can be well
 155 represented by empirically derived scaling laws (Horedt and Neukum, 1984).
 156 These generally take the form of a power law $D = aE^b$, where a and b are
 157 constants with b varying between 1/4 and 1/3 depending on experimental
 158 conditions. More complex relationships can also be derived, which include
 159 secondary effects such as target/impactor material properties and impact
 160 angle (Holsapple and Schmidt, 1982; Horedt and Neukum, 1984).

161 For our study we only have one observable to describe the meteorite
 162 impact with Mars - the crater diameter. Therefore, we only aim to determine
 163 one parameter - the impact energy E , which is the most important factor
 164 determining crater diameter. To this end we use the cratering database
 165 of Holsapple et al. (2003), which covers laboratory drop tests, high energy

166 impact studies, and craters formed by conventional and nuclear detonations
 167 prior to 1998. We augmented this database by including additional studies
 168 and recently published data for: explosions (Nordyke, 1962; Goto et al., 2001;
 169 Ambrosini et al., 2002); impacts (Schmidt and Housen, 1987); and drop tests
 170 (Barnouin-Jha et al., 2007). To fill in the higher energy regime we also
 171 include a study of volcanic caldera diameters (Sato and Taniguchi, 1997).

172 The resulting database contains information on crater diameters for a
 173 large range of source energies, types, and incidence angles with target ma-
 174 terials ranging from water and ice through to rocks and metals. From this
 175 database we selected experimental data for non-metal solids (rock, sand,
 176 coral, and alluvium) at standard Terrestrial conditions (1 bar pressure and
 177 1 g gravitational acceleration). Many of the test explosions were either buried
 178 at significant depth or detonated above ground. To ensure that the measure-
 179 ments were relevant to meteorite impacts we only included explosions where
 180 the ratio of source depth to final crater diameter was in the range -0.05 to
 181 0.2. These limits were chosen because typically impact craters have a depth
 182 of one fifth of their diameter, so an explosion with a depth-diameter ratio
 183 of 0.2 or less should still be representative of an impact, whereas the small
 184 negative value of -0.05 was chosen so as to include explosives that were deto-
 185 nated just above the surface for experimental reasons (on a short rig/mount
 186 for example), but gave crater sizes indistinguishable from surface explosions.
 187 The equivalent source depth for the volcanic calderas is unknown, but they
 188 appear to follow the same scaling as the higher energy explosions so we also
 189 included these points.

190 Figure 2 shows energy versus crater diameter for our subset of the database.

191 Data from selected individual studies are also plotted for comparison and
 192 to show the scatter that occurs even under relatively uniform experimental
 193 conditions. A standard unit of energy for many of the older studies was
 194 equivalent pounds of TNT (lbs TNT) for small explosions or kilo-ton yields
 195 for nuclear detonations. We have converted these units into Joules using 1 lb
 196 TNT = 1.90 MJ and 1 kilo-ton = 4.18 TJ (Shoemaker, 1983) and included
 197 them as a secondary axes for ease of comparison.

198 Despite energy variations over 10 orders of magnitude, the dependence of
 199 crater diameter on energy is remarkably linear - especially over the $D = 2$ –
 200 148 m range, which is our region of primary interest as it corresponds to the
 201 new craters detected from orbit by Malin et al. (2006). There is a discrepancy
 202 for very low impact energy drop tests, where crater forming processes appear
 203 more efficient. However, such low energy events are not relevant for our study
 204 as ablation by Mars' atmosphere prevents the formation of craters smaller
 205 than 0.3 m (Popova et al., 2003). Explosion and impact generated craters
 206 give consistent results, which is fortunate as it is not possible to perform
 207 controlled impact studies with energy much over 10^9 J and chemical/nuclear
 208 explosive analogues must be used.

209 To determine the scaling relation we fitted a straight line to the E - D data
 210 in log-log space using robust estimation to minimise the absolute deviation
 211 (L1-norm) between model and data (Press et al., 1992). This method is less
 212 susceptible to outliers than the more usual least squares L2-norm minimi-
 213 sation. Only measurements with energies greater than 10^5 J were used to
 214 create the line of best fit (corresponding to craters over 0.3 m diameter). Our

215 resulting scaling law for craters under Terrestrial conditions is given by:

$$D = 8.8_{-3.5}^{+2.6} \times 10^{-3} E^{0.32 \pm 0.01} \quad (1)$$

216 where D is in meters and E is in Joules. The error bars on the constant
 217 of proportionality were chosen such that the minimum/maximum range en-
 218 compassed 68% of the measured datapoints. The error bars on the power
 219 are formal 1- σ errors from the robust fit. Unknown factors relating to the
 220 impact conditions, such as impact angle and rock properties, are included in
 221 our overall error estimate, which provides a measure of the uncertainty in
 222 determining the impactor energy from crater diameter alone.

223 The gravity on Mars is 3.73 ms^{-2} compared to 9.81 ms^{-2} on Earth. Be-
 224 cause lower gravity makes craters easier to excavate, this will result in slightly
 225 larger craters than predicted by the above scaling law, which was derived for
 226 Terrestrial conditions. However, compared to other factors and scatter in
 227 the data, this effect is relatively small. Horedt and Neukum (1984) review
 228 the early literature and suggest that gravity affects the crater diameter by a
 229 factor of $(g_{\oplus}/g)^{3/16}$, where g and $g_{\oplus}$ are the surface gravity on the planet and
 230 on Earth respectively. This value is consistent with the impact of Ranger 8
 231 on the lunar surface, which resulted in a crater approximately 13 m across
 232 (Whitaker, 1972), compared to the expected 9.5 m from missile impacts with
 233 similar materials and trajectories at the White Sands missile range in Nevada
 234 (Moore and Baldwin, 1968). The Deep Impact crater may provide additional
 235 insight on gravity scaling under very low gravity conditions but at present
 236 the crater diameter is hard to determine (Busko et al., 2007).

237 Therefore, our overall scaling law for a general planet - including uncer-
 238 tainty for unknown source/target parameters and a gravity correction - is

239 given in SI units (D in meters, E in Joules, and g in ms^{-2}) by:

$$D = 8.8_{-3.5}^{+2.6} \times 10^{-3} E^{0.32 \pm 0.01} \left(\frac{g_{\oplus}}{g} \right)^{3/16} \quad (2)$$

240 This compares well with the diameters of artificial impact craters created
241 by Ranger 7–9 and Apollo 13 and 14 booster impacts measured from Apollo
242 16 (Whitaker, 1972).

243 2.3. Seismic Efficiency - k

244 Only a small fraction of a meteorite's kinetic energy goes into creating
245 seismic waves. The bulk of the energy is expended as heat during deformation
246 of the target along with the energy used to excavate the crater. The fraction
247 of the total impact energy E converted into seismic energy E_s is known as
248 the seismic efficiency k , which can be derived experimentally using impacts
249 or explosive analogues. Combining k with the cratering rate and impact
250 energy to crater diameter scaling law will allow us to determine the number
251 of seismic events with a given energy.

252 Compiled experimental data from laboratory impact tests (McGarr et al.,
253 1969), missiles impacts (Latham et al., 1970b), lunar module ascent stage
254 (LM) and the Apollo 13 third stage Saturn booster (S-IVB) lunar impacts
255 (Latham et al., 1970a), surface detonated nuclear explosions (Pomeroy, 1963),
256 and underground nuclear detonations (Patton and Walter, 1993) are shown
257 in Figure 3. The results of numerical calculations by Walker (2003) are also
258 shown for comparison. Note that underground nuclear explosions are much
259 more efficient at coupling seismic energy into the ground and have high seis-
260 mic efficiencies of around $0.2\text{--}4 \times 10^{-2}$. However, these efficiencies are not
261 representative of surface impacts.

262 The experimental data suggest a seismic efficiency in the range $k=10^{-4}$ –
 263 10^{-6} . For comparison, Chyba et al. (1998) adopted a value of $k=10^{-4}$ for the
 264 Earth, Davis (1993) used $k=10^{-6}$ for Mars, and Lognonne et al. (2009) used
 265 $k=10^{-5}$ in a recent study of lunar impacts. The value of $k=10^{-4}$ is closer to
 266 the values calculated by Walker (2003), however, these calculations show a
 267 trend for decreasing k as the impact energy increases, suggesting that lower
 268 values are appropriate for our energy range of interest ($E > 10^{11}$ J).

269 We adopt a value of $k=2 \times 10^{-5}$, in keeping with the high energy results.
 270 However, this value contains considerable uncertainty. The experimental
 271 conditions for the lower k values may also not be an accurate representation
 272 of what happens during a high energy impact event. For example, the value
 273 from Pomeroy (1963) was obtained from a 5 Mton nuclear device detonated
 274 on a shallow water barge over a Pacific atoll, which provided sub-optimal con-
 275 ditions for the creation of seismic waves. Therefore, our value of $k=2 \times 10^{-5}$
 276 most probably represents a conservative estimate and must include a large
 277 factor of 10 error.

278 *2.4. Relation Between Seismic Moment and Seismic Energy*

279 In order to calculate synthetic seismograms we require the equivalent
 280 seismic moment of the source. Therefore, we need to know how the seismic
 281 moment M is related to the seismic energy E_s . We can then use the seismic
 282 efficiency and impact energy to calculate M for a particular impact event.

283 Seismic energy is difficult to determine experimentally as attenuation in
 284 the Earth's crust preferentially removes high frequencies, which contain most
 285 of the energy. This has lead to some controversy in determining the relation
 286 between M and E_s , which is discussed by Shearer (2009).

287 Ide and Beroza (2001) compiled data from six studies: from micro-earthquakes
 288 to teleseismic events and concluded that the relation $E_s = 3 \times 10^{-5} M$ fits the
 289 data very well, although there is a large amount of scatter about this mean
 290 value (the bulk of the data lies in the range $E_s/M = 10^{-6}$ to 10^{-4}).

291 However, the attenuation problem means that seismic energy estimation
 292 from teleseismic events is very difficult, so the studies of small local earth-
 293 quakes may be more representative. They also cover the range of energies
 294 of interest for our study. Figure 4 shows a compilation of three large studies
 295 where seismic energy and moment were determined. A robust (L1 norm) fit
 296 to this dataset yields the relation:

$$E_s = 4.8_{-1.8}^{+2.9} \times 10^{-9} M^{1.24 \pm 0.01} \quad (3)$$

297 Error bars are 1- σ , E_s is in Joules, and M is in Nm. The results from
 298 the impact modelling by Walker (2003) (assuming $k = 2 \times 10^{-5}$) and nuclear
 299 explosions by Patton and Walter (1993) are also shown. These are in broad
 300 agreement with both trends. However, the $E_s = 4.8 \times 10^{-9} M^{1.24}$ relation
 301 provides a better overall fit to the data in this energy range and is the one
 302 we use here.

303 2.5. Overall Relation Between Crater Diameter and Seismic Moment

304 To summarise the results of the previous sections, we have considered
 305 the following key variables: impact energy E in Joules; crater diameter D in
 306 meters; dimensionless seismic efficiency k ; seismic energy E_s in Joules; and
 307 seismic moment M in Nm. These quantities are related by:

$$D = a E^b \left(\frac{g_{\oplus}}{g} \right)^{3/16} \quad (4)$$

$$E_s = kE \quad (5)$$

$$E_s = cM^d \quad (6)$$

Our observed quantity is D and we require the equivalent seismic moment M , which is given by the overall relation:

$$M(D) = \left(\frac{D}{a}\right)^{\frac{1}{bd}} \left(\frac{k}{c}\right)^{\frac{1}{d}} \left(\frac{g}{g_{\oplus}}\right)^{\frac{3}{16bd}} \quad (7)$$

The values of the constants and their fractional errors are given in Table 1. This table also gives the contribution of each of the uncertainties to the overall fractional error in M for a given observed D using the error propagation equations in Bevington and Robinson (1992). The overall fractional error in our estimate of M is of order $\sigma_M/M \approx 8$. Therefore, there is around an order of magnitude uncertainty in seismic moment for a given impact event.

Table 2 summarises our results for the impact induced seismicity of Mars and gives the number of impact events with a given D and M in each decadal energy bin, with the impact rates determined from the modified Hartmann (2005) isochrons. The seismic moment magnitude M_w , defined by $M_w = 2/3(\log_{10} M - 9.1)$ (Kanamori, 1977), is also given for ease of comparison with terrestrial events.

3. Mars Interior Model

Current knowledge of Mars' interior is reviewed by Lognonne (2005). Here we are concerned with velocity, density, and seismic attenuation as these determine how seismic waves propagate throughout the interior.

326 3.1. Velocity Structure

327 Early models of Mars' interior are summarised by Okal and Anderson
 328 (1978). More recently, Sohl and Spohn (1997) presented two end-member
 329 models of Mars' velocity and density structure, such that either the fit to SNC
 330 meteorite bulk composition or the observed moment of inertia was optimised.
 331 As a representative model, we adopt model A from Sohl and Spohn (1997)
 332 (optimised to fit the observed moment of inertia), which is plotted in Figure 5.

333 3.1.1. Seismic Attenuation: Q

334 Seismic attenuation is parameterised by the quality factor Q . Obviously,
 335 this is a significant unknown at seismic frequencies for Mars. However, it
 336 can be estimated at long periods (~ 20000 s) from measurements of the secu-
 337 lar acceleration of Phobos and extrapolated to seismic frequencies. Zharkov
 338 and Gudkova (1997) determined the mean dissipative factors at tidal fre-
 339 quencies, and using a five-layer density model and uniform Q model inferred
 340 values of $Q_s=177$ assuming a solid core and $Q_s=261$ assuming a liquid core
 341 for periods of 1s. Yoder et al. (2003) show Mars has a liquid core, so we
 342 assume $Q_s=261$ throughout Mars' interior. Q_p is assumed to be $2.25Q_s$ fol-
 343 lowing standard relations (e.g. Shearer, 2009). This yields an attenuation
 344 value which is similar to the average for the upper 2000 km Earth's mantle
 345 (where $Q_s=280$). Without structural information we cannot constrain depth
 346 dependence. However, the attenuation is a path-integral effect so a uniform
 347 value for the Martian mantle is sufficient for our purposes.

348 4. Seismic Source Function

349 The seismic frequency content of the source will depend on the impact en-
 350 ergy. Larger events will have more energy concentrated at lower frequencies.
 351 Here, we assume a simple displacement source function $S(f)$ after Boatwright
 352 (1980) as used by Abercrombie (1995).

$$\frac{S(f)}{S_0} = \frac{1}{(1 + (f/f_c)^\gamma)^{1/\gamma}} \quad (8)$$

353 where f is frequency, f_c is the corner frequency, S_0 is a reference displace-
 354 ment at long period, $\gamma=2$, and $n=2$. This model gives a source spectra that
 355 falls in between those for an idealised fault (Brune, 1970), which has slightly
 356 less high frequency content, and that for an underground nuclear explosion
 357 (Mueller and Murphy, 1971), which has slightly higher frequency content.
 358 All three models tend to a classic f^{-2} dependence at high frequency.

359 The corner frequency f_c has a weak dependence on the source moment M ,
 360 such that $f_c \propto M^{-1/3}$ (Shearer, 2009). For micro-earthquakes the relation is
 361 given approximately by:

$$f_c = 2 \times 10^5 M^{-1/3} \quad (9)$$

362 which was derived using the corner frequencies determined from high sample
 363 rate measurements of micro-earthquakes by Abercrombie (1995). Figure 6
 364 shows the source spectra using Equations 8 and 9 for a range of seismic
 365 moments.

366 Note that for all but the largest events, the source spectrum is flat until
 367 relatively high frequencies and will have limited effect on the seismogram

frequency content compared to seismic attenuation in the crust and mantle.
Therefore, for most impacts a delta function in displacement would provide
an excellent approximation to the source.

5. Synthetic Seismograms

Synthetic Martian seismograms were generated using the direct solution
method (DSM) (Geller and Ohminato, 1994; Geller and Takeuchi, 1995;
Takeuchi et al., 1996), which allows generation of accurate full waveform
synthetics at global scales.

During high energy/high velocity impacts, such as a meteorite impacting
a planet, the impactor tends to fragment and vapourise. This leads to the
creation of radially symmetric craters for all but the steepest impact angles.
Therefore, the most appropriate seismic source is a purely isotropic (explosive)
source (Stein and Wysession, 2002), defined by the moment tensor:

$$\mathbf{M}_T = \begin{pmatrix} M_{11} & 0 & 0 \\ 0 & M_{22} & 0 \\ 0 & 0 & M_{33} \end{pmatrix} \quad (10)$$

where $M_{11} = M_{22} = M_{33}$. The scalar seismic moment M is then defined by:

$$M = \frac{1}{\sqrt{2}} \left(\sum_{i=1}^3 \sum_{j=1}^3 M_{ij}^2 \right)^{\frac{1}{2}} \quad (11)$$

which reduces to $M = \sqrt{3/2}M_{11}$ in our case.

Modelling shallow events is too computationally expensive (Kawai et al.,
2006), but is only required for surface wave generation. In this study we
are only interested in body waves amplitudes, as these waves sound the deep

interior. Body wave amplitudes are affected very little by changing the model source depth (Figure 7), so for our purposes accurate impact-generated body waves can be modelled using a shallow source at 50 km depth.

Models were run at 1, 2, 4, and 8 Hz maximum frequency using a 64000 gridpoint vertical velocity model and spherical harmonics up to degree 64000 to give a model spatial resolution of around 50 m, which gave 10 times oversampling of the shortest wavelength seismic waves. The 8 Hz synthetics took 48khr equivalent CPU time. Computation time scaled as the cube of frequency, so it was not computationally practical to model frequencies higher than 8 Hz. Although the DSM method calculates the full wavefield, in the following we only analyse P waves. These are expected to be the most energetic phase for non-shear sources like impacts and so represent the best chance for probing the interior.

Figure 8 shows waveforms calculated for each maximum modelled frequency for three source-receiver offsets. If waveforms are not modelled with a high enough maximum frequency then “ringing” effects will be seen around the arrivals and the amplitudes of the waveforms will not have converged. Comparison of synthetics with different maximum frequencies was used to check for amplitude convergence. Due to seismic attenuation, it was not necessary to model frequencies higher than 8 Hz to accurately model synthetic seismograms with offsets over 30° . However, closer seismograms travel shorter distances and have been attenuated less. Therefore, at close range higher maximum frequency is required. These effects are just visible in the 20° offset synthetics in Figure 8, which have not quite fully converged. Therefore, amplitudes at offsets less than 20° will be underestimated, but as we are

411 primarily concerned with teleseismic events (as these provide information on
412 the deep structure) this does not pose a problem for our analysis.

413 The seismic displacement source function from Section 4 was applied
414 to the synthetics although this had negligible effect compared to seismic
415 attenuation for events with $M \leq 10^{14}$ Nm.

416 Figure 9 shows a record section obtained from our modelling. Travel time
417 curves of the various phases were calculated by ray tracing using the Tau-p
418 toolkit (Crotwell et al., 1999) and are overlain for comparison. On Mars the
419 P-wave shadow zone starts at around 110° offset.

420 Figure 10 shows the frequency content of the P-wave envelope as a func-
421 tion of source-receiver offset angle. Seismic attenuation limits the frequency
422 content of teleseismic arrivals within the 0.4–4 Hz band. More distant ar-
423 rivals are attenuated more and have a lower frequency content. Note that
424 for Earth, accurate seismograms can be obtained using a maximum modelled
425 frequency of 2 Hz (Kawai et al., 2006). However, the smaller radius of Mars
426 requires modelling to higher frequencies (4–8 Hz) as there is less distance for
427 the seismic waves to be attenuated and the frequency content is thus higher.

428 6. Results

429 Figure 11 shows the maximum P-wave amplitude as a function of epicen-
430 tral distance for a range of impact equivalent source moments after band pass
431 filtering with a 0.4–4 Hz 16 pole Butterworth filter. The difference between
432 the 1, 2, 4, and 8 Hz maximum frequency models again shows that for close
433 events with offsets less than 20° the amplitudes have not quite fully converged
434 and should be regarded as lower limits. These predicted amplitudes can be

435 compared to expected background noise levels and instrument sensitivities
 436 to determine how many events would be likely to be detected per year.

437 The major noise source on Mars is likely to be induced by winds. Lognonne
 438 et al. (1996) calculated a conservative estimate of the wind induced noise
 439 spectral density of $10^{-9} \text{ ms}^{-2}\text{Hz}^{-1/2}$. For comparison, on Earth the noise
 440 spectral density in our frequency range of interest (0.4-4 Hz) is around $3 \times 10^{-9} \text{ ms}^{-2}\text{Hz}^{-1/2}$
 441 for a low noise site if the ocean wave-induced microseismic noise peak at
 442 0.2 Hz is ignored (Peterson, 1993). Mars should be much quieter than the
 443 Earth and a noise spectral density of $10^{-10} \text{ ms}^{-2}\text{Hz}^{-1/2}$ is not unreason-
 444 able for a wind-shielded instrument or during calm meteorological periods
 445 (e.g. night time).

446 The instrumental noise of typical terrestrial broad band seismometers
 447 (Nanometrics Trillium240, Guralp CMG3T, and Streckeisen STS2) range
 448 from $0.5\text{--}5 \times 10^{-10} \text{ ms}^{-2}\text{Hz}^{-1/2}$ at 1 Hz. For Mars, a very broad band seis-
 449 mometer developed at IPGP in France has a noise level approaching $10^{-10} \text{ ms}^{-2}\text{Hz}^{-1/2}$
 450 at 1 Hz (Lognonne et al., 2000). Another approach would be to use an ex-
 451 tended frequency range micro-seismometer, such as the micro-machined sili-
 452 con instrument under development for planetary applications by Pike et al.
 453 (2005). This seismometer is predicted to have a noise level of around 1--
 454 $3 \times 10^{-8} \text{ ms}^{-2}\text{Hz}^{-1/2}$ at 1 Hz, although further improvements in performance
 455 may be possible.

456 Therefore, to cover the range of likely background and instrument noise
 457 levels we consider three rms noise spectral densities: $10^{-10} \text{ ms}^{-2}\text{Hz}^{-1/2}$,
 458 $10^{-9} \text{ ms}^{-2}\text{Hz}^{-1/2}$, and $10^{-8} \text{ ms}^{-2}\text{Hz}^{-1/2}$. The lowest noise would corre-
 459 spond to a high specification broad band instrument under quiet condi-

tions, whereas the highest noise case represents an exploration-class instrument in a non-ideal deployment. These noise spectral densities n_{rms} were converted into maximum peak acceleration a_{peak} using the formula $a_{peak} = 1.25n_{rms}\sqrt{(f_2 - f_1)}$, where f_1 – f_2 is the frequency range under consideration (Havskov and Alguacil, 2004).

Under nominal noise conditions ($10^{-9} \text{ ms}^{-2}\text{Hz}^{-1/2}$) Figure 11 shows that to detect an event at teleseismic distances (offset $\Delta \geq 60^\circ$) requires a seismic moment over 10^{13} Nm ($M_w=2.6$), or equivalently an impact producing a 100 m diameter or larger crater. From Table 2 such large craters are produced every 1–8 years or 4 years on average. A globally detectable event would require a moment of around 10^{14} Nm ($M_w=3.3$).

Figure 12 summarises the number of detectable impacts for each noise level in each of the decadal energy bins in Table 2. The number of detectable impacts at a given site per year per energy bin N_e is given by:

$$N_e = \frac{(1 - \cos \theta)}{2} i_e \quad (12)$$

where i_e is the number of impacts per year over the whole of Mars in each energy bin, θ is the maximum offset at which an event is detectable, and the $(1 - \cos \theta)/2$ factor gives the fractional surface area of Mars over which an event would be detectable. The nominal noise case represents the most likely scenario and shows that a single seismic station would detect impacts with $\Delta \geq 60^\circ$ approximately once every 10 years. For the low noise case, which assumes a more sensitive instrument, around one such event occurs per year.

481 7. Discussion and Conclusions

482 In this paper we derive a relation between crater diameter and equiva-
 483 lent seismic moment for meteorite impacts. This is important for accurate
 484 modelling of the seismic signals produced by impacts. When combined with
 485 the impact rate determined from crater counting and recently observed new
 486 craters, synthetic seismograms can be modelled using a representative Mars
 487 internal model and used to predict the number of seismically detectable im-
 488 pacts per year.

489 Our seismic modelling shows that a detectable event requires a seismic
 490 moment of at least 10^{13} Nm to be detected above the predicted noise level
 491 at teleseismic distances ($\Delta \geq 60^\circ$) and a moment of 10^{14} Nm to be detected
 492 globally. These figures are consistent with a previous study by Mocquet
 493 (1999).

494 Our study suggests that globally detectable impact events on Mars are
 495 very rare and most probably do not provide a dependable way of probing
 496 the deep interior. Nominal assumptions about noise and seismic parameters
 497 suggest only one impact at $\Delta \geq 60^\circ$ is detectable every 10 years. If low noise
 498 conditions are assumed, detection of one such event per year is predicted. If
 499 in addition to low noise, we are very optimistic and use the more favourable
 500 end-members of the likely seismic parameter space, detections for offsets
 501 $\Delta \geq 60^\circ$ could be as high as 10 per year per site. However, based on current
 502 understanding of impact processes we regard this as unlikely.

503 Therefore, our estimates of meteorite detectability are much less than
 504 those of Davis (1993), who applied scaling to the lunar results and predicted
 505 around 20 globally detectable impact events per year. The main reason for

506 this discrepancy is the downward revision of the Martian impactor flux over
 507 the last decade or so. Comparing our Table 2 to Table III of Davis (1993)
 508 shows that we predict about two orders of magnitude less events for a given
 509 energy.

510 The seismicity due to impactors can be compared to that from shallow
 511 surface and lithospheric faulting estimated from faults observed on Mars'
 512 surface with high resolution imagery and topography measurements. By
 513 summing the product of the impact equivalent seismic moments and impact
 514 rates in Table 2, we estimate the total seismic moment release per year due to
 515 impacts to be 10^{13} – 10^{14} Nm. Golombek et al. (1992) estimated that around
 516 50 globally detectable ($M > 10^{14}$ Nm) faulting events should occur per year,
 517 with a total yearly moment release of 10^{18} Nm. A more recent analysis
 518 by Knapmeyer et al. (2006) predicts a yearly moment release from faults
 519 in the range 3×10^{16} – 5×10^{18} Nm, with around 100 $M > 10^{14}$ Nm events
 520 (depending on the fault model used). Therefore, these studies suggest that
 521 faulting induced seismicity should be a much more effective way to probe
 522 the Martian interior than meteorite impacts. However, this leaves us with
 523 the problem of locating the events, which would require measurements at
 524 multiple stations.

525 This paper shows that seismic waves generated by meteorite impacts
 526 probably cannot be relied upon for sounding the interior of Mars. If low
 527 noise and optimistic seismic parameters are assumed around 10 teleseismic
 528 range impacts could be detectable per year. However, using our nominal
 529 noise and parameter values predicts only one detectable teleseismic range
 530 impact every 10 years. Seismic efficiency and background noise levels are

the most poorly understood of our seismic parameters and introduce considerable uncertainty into the detection rates. Due to the large uncertainty, impacts should not be ruled out as a means for probing Mars' interior, but are expected to be of secondary importance to faulting. Local impacts will be more detectable and occur more often - providing valuable constraints on near-surface properties - but are less well suited to probing the deep interior. However, if even one global impact event could be seismically detected and the crater identified, this would provide some of the best constraints we have so far on Mars' interior structure.

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References

- Abercrombie, R. E., 1995. Earthquake source scaling relationships from -1 to 5 M_L using seismograms recorded at 2.5-km depth. *J. Geophys. Res.* B100, 24015–24036.
- Acuna, M. H., Connerney, J. E. P., Ness, N. F., Lin, R. P., Mitchell, D., Carlson, C. W., McFadden, J., Anderson, K. A., Reme, H., Mazelle, C., Vignes, D., Wasilewski, P., Cloutier, P., 1999. Global distribution of crustal magnetization discovered by the Mars Global Surveyor MAG/ER experiment. *Science* 284, 790–793.

- 553 Ambrosini, R. D., Luccioni, B. M., Danesi, R. F., Riera, J. D., Rocha, M. M.,
554 2002. Size of craters produced by explosive charges on or above the ground
555 surface. *Shock Waves* 12, 69–78.
- 556 Anderson, D. L., Duennebier, F. K., Latham, G. V., Toksoz, M. F., Kovach,
557 R. L., Knight, T. C. D., Lazarewicz, A. R., Miller, W. F., Nakamura, Y.,
558 Sutton, G., 1976. Viking seismic experiment. *Science* 194, 1318–1321.
- 559 Barnouin-Jha, O. S., Yamamoto, S., Toriumi, T., Sugita, S., Matsui, T.,
560 2007. Non-intrusive measurements of crater growth. *Icarus* 188, 506–521.
- 561 Bevington, P. R., Robinson, D. K., 1992. Data reduction and error analysis
562 for the physical sciences, 2nd Edition. WBC/McGraw-Hill, New York.
- 563 Bills, B. G., Ferrari, A. J., 1977. Lunar density model consistent with topo-
564 graphic, gravitational, librational, and seismic data. *J. Geophys. Res.* 82,
565 1306–1314.
- 566 Boatwright, J., 1980. A spectral theory for circular seismic sources - simple
567 estimates of source dimension, dynamic stress drop, and radiated seismic
568 energy. *Bull. Seism. Soc. Am.* 70, 1–27.
- 569 Brune, J. N., 1970. Tectonic stress and spectra of seismic shear waves from
570 earthquakes. *J. Geophys. Res.* 75, 4997–5009.
- 571 Busko, I., Lindler, D., A'Hearn, M. F., White, R. L., 2007. Searching for
572 the Deep Impact crater on comet 9P/Tempel 1 using image processing
573 techniques. *Icarus* 187, 56–68.

- 574 Chyba, C. F., van der Vink, G. E., Hennet, C. B., 1998. Monitoring the com-
575 prehensive test ban treaty: Possible ambiguities due to meteorite impacts.
576 *Geophys. Res. Lett.* 25, 191–194.
- 577 Connerney, J. E. P., Acuna, M. H., Wasilewski, P. J., Ness, N. F., Reme, H.,
578 Mazelle, C., Vignes, D., Lin, R. P., Mitchell, D. L., Cloutier, P. A., 1999.
579 Magnetic lineations in the ancient crust of Mars. *Science* 284, 794–798.
- 580 Crotwell, H. P., Owens, T. J., Ritsema, J., 1999. The TauP Toolkit: Flexible
581 seismic travel-time and ray-path utilities. *Seis. Res. Lett.* 70, 154–160.
- 582 Davis, P. M., 1993. Meteoroid impacts as seismic sources on Mars. *Icarus*
583 105, 469–478.
- 584 Dziewonski, A. M., Anderson, D. L., 1981. Preliminary reference earth model.
585 *Phys. Earth Planet. Int.* 25, 297–356.
- 586 Gagnepain-Beyneix, J., Lognonne, P., Chenet, H., Lombardi, D., Spohn, T.,
587 2006. A seismic model of the Lunar mantle and constraints on temperature
588 and mineralogy. *Phys. Earth Planet. Int.* 159, 140–166.
- 589 Geller, R. J., Ohminato, T., 1994. Computation of synthetic seismograms and
590 their partial derivatives for heterogeneous media with arbitrary natural
591 boundary-conditions using the direct solution method. *Geophys. J. Int.*
592 116, 421–446.
- 593 Geller, R. J., Takeuchi, N., 1995. A new method for computing highly accu-
594 rate DSM synthetic seismograms. *Geophys. J. Int.* 123, 449–470.

- 595 Goins, N. R., Dainty, A. M., Toksoz, M. N., 1981. Lunar seismology - the
596 internal structure of the moon. *J. Geophys. Res.* B86, 5061–5074.
- 597 Goins, N. R., Lazarewicz, A. R., 1979. Martian seismicity. *Geophys. Res.*
598 *Lett.* 6, 368–370.
- 599 Golombek, M. P., Banerdt, W. B., Tanaka, K. L., Tralli, D. M., 1992. A
600 prediction of Mars seismicity from surface faulting. *Science* 258, 979–981.
- 601 Goto, A., Taniguchi, H., Yoshida, M., Ohba, T., Oshima, H., 2001. Effects
602 of explosion energy and depth to the formation of blast wave and crater:
603 Field explosion experiment for the understanding of volcanic explosion.
604 *Geophys. Res. Lett.* 28, 4287–4290.
- 605 Hartmann, W. K., 2005. Martian cratering 8: Isochron refinement and the
606 chronology of Mars. *Icarus* 174, 294–320.
- 607 Hartmann, W. K., 2007. Martian cratering 9: Toward resolution of the con-
608 troversy about small craters. *Icarus* 189, 274–278.
- 609 Havskov, J., Alguacil, G., 2004. *Instrumentation in Earthquake Seismology*.
610 Springer, Netherlands.
- 611 Holsapple, K. A., Housen, K. R., 2007. A crater and its ejecta: An interpre-
612 tation of Deep Impact. *Icarus* 187, 345–356.
- 613 Holsapple, K. A., Housen, K. R., Schmidt, R. M., Wauchope, C.,
614 2003. Explosive and impact crater database. (available online at
615 <http://keith.aa.washington.edu/craterdata/>).

- 616 Holsapple, K. A., Schmidt, R. M., 1982. On the scaling of crater dimensions
617 .2. impact processes. *J. Geophys. Res.* 87, 1849–1870.
- 618 Horedt, G. P., Neukum, G., 1984. Comparison of 6 crater-scaling laws. *Earth*
619 *Moon And Planets* 31, 265–269.
- 620 Ide, S., Beroza, G. C., 2001. Does apparent stress vary with earthquake size?
621 *Geophys. Res. Lett.* 28, 3349–3352.
- 622 Kanamori, H., 1977. Energy-release in great earthquakes. *J. Geophys. Res.*
623 82, 2981–2987.
- 624 Kawai, K., Takeuchi, N., Geller, R. J., 2006. Complete synthetic seismograms
625 up to 2 Hz for transversely isotropic spherically symmetric media. *Geophys.*
626 *J. Int.* 164, 411–424.
- 627 Khan, A., Connolly, J. A. D., MacLennan, J., Mosegaard, K., 2007. Joint
628 inversion of seismic and gravity data for lunar composition and thermal
629 state. *Geophys. J. Int.* 168, 243–258.
- 630 Khan, A., Mosegaard, K., Rasmussen, K. L., 2000. A new seismic velocity
631 model for the Moon from a Monte Carlo inversion of the Apollo lunar
632 seismic data. *Geophys. Res. Lett.* 27, 1591–1594.
- 633 Knapmeyer, M., Oberst, J., Hauber, E., Wahlisch, M., Deuchler, C., Wag-
634 ner, R., 2006. Working models for spatial distribution and level of Mars’
635 seismicity. *J. Geophys. Res.* 111, E11006.
- 636 Latham, G., Ewing, M., Dorman, J., Press, F., Toksoz, N., Sutton, G.,

- 637 Meissner, R., Duennebi, F., Nakamura, Y., Kovach, R., Yates, M., 1970a.
638 Seismic data from man-made impacts on the moon. *Science* 170, 620–626.
- 639 Latham, G. V., McDonald, W. G., Moore, H. J., 1970b. Missile impacts as
640 sources of seismic energy on the moon. *Science* 168, 242–245.
- 641 Lognonne, P., 2005. Planetary seismology. *Annual Review Of Earth And*
642 *Planetary Sciences* 33, 571–604.
- 643 Lognonne, P., Beyneix, J. G., Banerdt, W. B., Cacho, S., Karczewski, J. F.,
644 Morand, M., 1996. Ultra broad band seismology on InterMarsNet. *Plan.*
645 *& Space Sci.* 44, 1237–1249.
- 646 Lognonne, P., Gagnepain-Beyneix, J., Chenet, H., 2003. A new seismic model
647 of the Moon: implications for structure, thermal evolution and formation
648 of the Moon. *Earth Planet. Sci. Let.* 211, 27–44.
- 649 Lognonne, P., Giardini, D., Banerdt, B., Gagnepain-Beyneix, J., Mocquet,
650 A., Spohn, T., Karczewski, J. F., Schibler, P., Cacho, S., Pike, W. T.,
651 Cavoit, C., Desautiez, A., Favade, M., Gabsi, T., Simoulin, L., Striebig,
652 N., Campillo, M., Deschamp, A., Hinderer, J., Leveque, J. J., Moatag-
653 ner, J. P., Rivera, L., Benz, W., Breuer, D., Defraigne, P., Dehant, V.,
654 Fujimura, A., Mizutani, H., Oberst, J., 2000. The NetLander very broad
655 band seismometer. *Plan. & Space Sci.* 48, 1289–1302.
- 656 Lognonne, P., Le Feuvre, M., Johnson, C. L., Weber, R. C., 2009. Moon
657 meteoritic seismic hum: Steady state prediction. *J. Geophys. Res.* 114,
658 E12003.

- 659 Malin, M. C., Edgett, K. S., Posiolova, L. V., McColley, S. M., Dobrea, E.
 660 Z. N., 2006. Present-day impact cratering rate and contemporary gully
 661 activity on Mars. *Science* 314, 1573–1577.
- 662 Mayeda, K., Walter, W. R., 1996. Moment, energy, stress drop, and source
 663 spectra of western United States earthquakes from regional coda envelopes.
 664 *J. Geophys. Res.* B101, 11195–11208.
- 665 McGarr, A., Latham, G. V., Gault, D. E., 1969. Meteoroid impacts as sources
 666 of seismicity on the Moon. *J. Geophys. Res.* 74, 5981–5994.
- 667 Melosh, H. J., 1980. Cratering mechanics - observational, experimental, and
 668 theoretical. *Annual Review Of Earth And Planetary Sciences* 8, 65–93.
- 669 Melosh, H. J., 1989. Impact cratering - a geological process. Vol. 11 of Oxford
 670 Monographs on Geology and Geophysics. Oxford University Press.
- 671 Mocquet, A., 1999. A search for the minimum number of stations needed for
 672 seismic networking on Mars. *Plan. & Space Sci.* 47, 397–409.
- 673 Moore, H. J., Baldwin, R. B., 1968. Ranger VIII and gravity scaling of Lnar
 674 craters. *Science* 159, 333–334.
- 675 Mueller, R. A., Murphy, J. R., 1971. Seismic characteristics of underground
 676 nuclear detonations 1. seismic spectrum scaling. *Bull. Seism. Soc. Am.* 61,
 677 1675–1692.
- 678 Nakamura, Y., 1976. Seismic energy transmission in lunar-surface zone de-
 679 termined from signals generated by movement of lunar rovers. *Bull. Seism.*
 680 *Soc. Am.* 66, 593–606.

- 681 Nakamura, Y., 1983. Seismic velocity structure of the Lunar mantle. *J. Geo-*
 682 *phys. Res.* B88, 677–686.
- 683 Nakamura, Y., Koyama, J., 1982. Seismic-Q of the Lunar upper mantle. *J.*
 684 *Geophys. Res.* B87, 4855–4861.
- 685 Nordyke, M. D., 1962. An analysis of cratering data from desert alluvium. *J.*
 686 *Geophys. Res.* 67, 1965–1974.
- 687 Oberst, J., Nakamura, Y., 1987. Distinct meteoroid families identified on the
 688 Lunar seismograms. *J. Geophys. Res.* 92, E769–E773.
- 689 Okal, E. A., Anderson, D. L., 1978. Theoretical-models for Mars and their
 690 seismic properties. *Icarus* 33, 514–528.
- 691 Patton, H. J., Walter, W. R., 1993. Regional moment - magnitude relations
 692 for earthquakes and explosions. *Geophys. Res. Lett.* 20, 277–280.
- 693 Peterson, J., 1993. Observations and modeling of seismic background noise.
 694 *U.S. Geol. Surv. Tech. Rept.* 93-322, 1–95.
- 695 Pike, W. T., Standley, I. M., Banerdt, W. B., Mar. 2005. A high-sensitivity
 696 broad-band seismic sensor for shallow seismic sounding of the Lunar re-
 697 golith. In: S. Mackwell & E. Stansbery (Ed.), 36th Annual Lunar and
 698 Planetary Science Conference. Vol. 36 of Lunar and Planetary Inst. Tech-
 699 nical Report. p. 2002.
- 700 Pomeroy, P. W., 1963. Long period seismic waves from large, near-surface
 701 nuclear explosions. *Bull. Seism. Soc. Am.* 53, 109–149.

- 702 Popova, O., Nemtchinov, I., Hartmann, W. K., 2003. Bolides in the present
703 and past martian atmosphere and effects on cratering processes. *Meteorit-*
704 *ics & Planet. Sci* 38, 905–925.
- 705 Press, W. H., Flannery, B. P., Teukolsky, S. A., Vetterling, W. T., 1992.
706 *Numerical Recipes*, 2nd Edition. Cambridge Univ. Press, Cambridge UK.
- 707 Prieto, G. A., Shearer, P. M., Vernon, F. L., Kilb, D., 2004. Earthquake
708 source scaling and self-similarity estimation from stacking P and S spectra.
709 *J. Geophys. Res.* B109, B08310.
- 710 Quantin, C., Mangold, N., Hartmann, W. K., Allemand, P., 2007. Possible
711 long-term decline in impact rates - 1. Martian geological data. *Icarus* 186,
712 1–10.
- 713 Richardson, J. E., Melosh, H. J., Artemeiva, N. A., Pierazzo, E., 2005. Impact
714 cratering theory and modeling for the Deep Impact mission: From mission
715 planning to data analysis. *Space Sci. Rev.* 117, 241–267.
- 716 Sato, H., Taniguchi, H., 1997. Relationship between crater size and ejecta
717 volume of recent magmatic and phreato-magmatic eruptions: Implications
718 for energy partitioning. *Geophys. Res. Lett.* 24, 205–208.
- 719 Schmidt, R. M., Housen, K. R., 1987. Some recent advances in the scaling of
720 impact and explosion cratering. *Int. J. Impact Eng.* 5, 543–560.
- 721 Shearer, P. M., 2009. *Introduction to Seismology*, 2nd Edition. Cam. Univ.
722 Press, Cambridge.

- 723 Shoemaker, E. M., 1983. Asteroid and comet bombardment of the earth.
724 Annual Review Of Earth And Planetary Sciences 11, 461–494.
- 725 Sleep, N. H., 1994. Martian plate tectonics. J. Geophys. Res. E99, 5639–5655.
- 726 Sohl, F., Schubert, G., Spohn, T., 2005. Geophysical constraints on the
727 composition and structure of the Martian interior. J. Geophys. Res. 110,
728 E12008.
- 729 Sohl, F., Spohn, T., 1997. The interior structure of Mars: Implications from
730 SNC meteorites. J. Geophys. Res. E102, 1613–1635.
- 731 Stein, S., Wysession, M., 2002. An Introduction to Seismology, Earthquakes
732 and Earth Structure. Wiley-Blackwell.
- 733 Stevenson, D. J., 2001. Mars' core and magnetism. Nature 412, 214–219.
- 734 Takeuchi, N., Geller, R. J., Cummins, P. R., 1996. Highly accurate P-SV
735 complete synthetic seismograms using modified DSM operators. Geophys.
736 Res. Lett. 23, 1175–1178.
- 737 Walker, J. D., 2003. Loading sources for seismological investigations of as-
738 teroids and comets. Int. J. Impact Eng. 29, 757–769.
- 739 Whitaker, E. A., 1972. Photogeology: Part I. Artificial lunar impact craters
740 photographed by Apollo 16. Apollo 16 Preliminary Science Report NASA
741 SP-315, 29.39–29.44.
- 742 Yoder, C. F., Konopliv, A. S., Yuan, D. N., Standish, E. M., Folkner, W. M.,
743 2003. Fluid core size of Mars from detection of the solar tide. Science 300,
744 299–303.

745 Zharkov, V. N., Gudkova, T. V., 1997. On the dissipative factor of the Mar-
746 tian interiors. *Plan. & Space Sci.* 45, 401–407.

Quantity	Value	Fractional error		Contribution to fractional error σ_M/M		
$\mathbf{D} = aE^b$						
a	8.8×10^{-3}	$\frac{\sigma_a}{a}$	=	0.35	$\frac{1}{bd} \frac{\sigma_a}{a}$	= 0.9
b	0.32	$\frac{\sigma_b}{b}$	=	0.03	$\frac{1}{b^2d} \ln \left[\frac{D}{a} \left(\frac{g}{g_{\oplus}} \right)^{\frac{3}{16}} \right] \frac{\sigma_b}{b}$	= 2.0†
$\mathbf{E}_s = kE$						
k	2×10^{-5}	$\frac{\sigma_k}{k}$	=	10	$\frac{1}{d} \frac{\sigma_k}{k}$	= 8.0
$\mathbf{E}_s = cM^d$						
c	4.8×10^{-9}	$\frac{\sigma_c}{c}$	=	0.49	$\frac{1}{d} \frac{\sigma_c}{c}$	= 0.4
d	1.24	$\frac{\sigma_d}{d}$	=	0.01	$\frac{1}{d^2} \ln \left[\left(\frac{D}{a} \left(\frac{g}{g_{\oplus}} \right)^{\frac{3}{16}} \right)^{\frac{1}{b}} \frac{k}{c} \right] \frac{\sigma_d}{d}$	= 0.2†
Total σ_M/M						8.3

Table 1: Contribution to the $M(D)$ error budget from the different derived quantities used in this paper. Error expressions were derived from the error propagation formulae in Bevington and Robinson (1992). The total fractional error σ_M/M , obtained by assuming independence of each error source and summing the variances, is of order 8 (around one order of magnitude in M for a given D). The error in seismic efficiency is the largest contributor to the overall error budget. † Calculated assuming a representative value of $D=50$ m.

$E_{\min}$ (J)	$E_{\max}$ (J)	M (Nm)	M (Nm)	M_w	Diameter (m)	Diameter (m)	N (per year)	$N_{\min}$ (per year)	$N_{\max}$ (per year)
10^8	10^9	2.3×10^9	1.5×10^{10}	0.2	0.7	3.67	7.64	300	1240
10^9	10^{10}	1.5×10^{10}	9.3×10^{10}	0.7	1.2	7.64	15.8	65.5	316
10^{10}	10^{11}	9.3×10^{10}	6.0×10^{11}	1.2	1.8	15.8	33.0	9.15	70.6
10^{11}	10^{12}	6.0×10^{11}	3.8×10^{12}	1.8	2.3	33.0	68.6	1.09	9.98
10^{12}	10^{13}	3.8×10^{12}	2.4×10^{13}	2.3	2.9	68.6	143	0.130	1.19
10^{13}	10^{14}	2.4×10^{13}	1.6×10^{14}	2.9	3.4	143	297	0.0140	0.142
10^{14}	10^{15}	1.6×10^{14}	1.0×10^{15}	3.4	3.9	297	617	0.00298	0.0153
10^{15}	10^{16}	1.0×10^{15}	6.4×10^{15}	3.9	4.5	617	1280	0.000220	0.00130
10^{16}	10^{17}	6.4×10^{15}	4.1×10^{16}	4.5	5.0	1280	2660	0.000018	0.000094
10^{17}	10^{18}	4.1×10^{16}	2.6×10^{17}	5.0	5.5	2660	5540	0.0000042	0.0000106

Table 2: Number of impact events N expected in each energy range $E_{\min}$ – $E_{\max}$. $N_{\min}$ and $N_{\max}$ give the minimum and maximum number of events based on uncertainties in the $D(E)$ relation. Equivalent seismic moments M and moment magnitudes M_w are calculated using the relations given in the main text.

Figure 1: Cratering 1-year isochron for Mars extrapolated from the Hartmann (2005) 3 Gyr isochron compared to measurements of new craters observed by Malin et al. (2006). The observed recent cratering suggests a correction factor of 1/3 must be applied to the Hartmann (2005) isochron for recent small craters. The bins defining the isochrons are the same as those used by Hartmann (2005), i.e. they give the number of craters per year in diameter intervals that are evenly spaced in $\log_{10}(\text{diameter})$ by $0.5 \log_{10} 2$ (e.g. 10–14.1 m, 14.1–20 m, 20–28.3 m, etc). Open/solid symbols contain single/multiple crater measurements respectively.

Figure 2: Measurements of crater diameter as a function of impact energy taken from the cratering database (CDB) of Holsapple et al. (2003) augmented with additional and post-1998 measurements. The solid and dashed lines show our derived scaling law and uncertainty bounds. Low energy ($E < 10^5$ J) impacts/explosions and drop tests are not representative of the meteorite impacts considered in this study and were not included when calculating the scaling law.

Figure 3: Seismic efficiency as a function of source energy compiled from various studies. The data are very scattered, indicating the large uncertainty in k . The dashed line shows our adopted value of $k=2 \times 10^{-5}$, which is broadly consistent (within an order of magnitude) with most of the experimental data. The k values for underground nuclear explosions are much higher and are not representative of impact processes.

Figure 4: Relationship between seismic moment and total radiated seismic energy. Data are from regional studies of small earthquakes (Abercrombie, 1995; Mayeda and Walter, 1996; Prieto et al., 2004), impact modelling (Walker, 2003), and nuclear explosions (Patton and Walter, 1993). The trend lines are shown from Ide and Beroza (2001) (dashed line) and this study (solid line). The $E_s = 4.8 \times 10^{-9} M^{1.24}$ trend fits these regional results the best and is considered most relevant to the present study.

Figure 5: Seismic velocity and density model for Mars from Sohl and Spohn (1997) (Model A).

Figure 6: Seismic displacement source spectra used for creation of the synthetic seismograms scaled by the seismic moment. The equivalent impact energy E is given for each case. Source spectra are essentially flat below 4 Hz for all but largest events. Hence, at teleseismic distances the high frequency content will be determined primarily by attenuation along the ray-path for all but the largest impacts. Spots and dashed line show the corner frequencies f_c which follow a f^{-3} trend. Frequency fall off above f_c follows a standard f^{-2} law.

Figure 7: Synthetic seismograms for a $M=10^{14}$ Nm event generated with 8 Hz maximum frequency and source depths of 50, 75, and 100km for a source-receiver offset of 60° . The difference between body wave amplitudes (P-waves in this plot) for different source depths is negligible and can be accounted for by slight differences in geometrical spreading. Therefore, while it is not computationally feasible to model surface events using DSM, a 50km source depth will give representative body wave amplitudes.

Figure 8: Waveforms for 1, 2, 4, and 8 Hz maximum modelled frequencies at source-receiver offsets Δ of 20, 40 and 60° for an event with $M=10^{14}$ Nm. Note that more distant events have lower frequency content due to increased seismic attenuation. At $\Delta=20^\circ$ some slight ringing is still visible in the 8 Hz model, implying that the synthetics have not quite converged at these small offsets.

Figure 9: Record section compiled from synthetic seismograms over a full Martian hemisphere. Travel time curves calculated using the Tau-P toolkit (Crotwell et al., 1999) for various phases generated using an isotropic source are also shown. A P-wave shadow zone exists beyond $\Delta=110^\circ$.

Figure 10: Spectra of P-wave envelope for a range of source-receiver offsets Δ for an event with $M=10^{14}$ Nm. Larger offsets have less high frequency content due to seismic attenuation. Most of the frequency content is in the 0.4–4 Hz range.

Figure 11: Maximum P-wave amplitude in the frequency range 0.4-4 Hz as a function of source-receiver offset and seismic moment for 8 Hz synthetics. Results from 1, 2, and 4 Hz models are also shown to give an idea of the convergence. Representative noise levels are also shown (dashed lines) over the same spectral range for noise spectral densities of $10^{-10} \text{ ms}^{-2}\text{Hz}^{-1/2}$ (low noise), $10^{-9} \text{ ms}^{-2}\text{Hz}^{-1/2}$ (nominal noise), and $10^{-8} \text{ ms}^{-2}\text{Hz}^{-1/2}$ (high noise). For the nominal noise case a globally detectable event must have a seismic moment of at around 10^{14} Nm ($M_w=3.3$), and at least 10^{13} Nm ($M_w=2.6$) to be detectable at 60° offset.

Figure 12: Detectability of meteorite impacts on Mars for deployments with low, nominal, and high noise levels. In the top panels, solid lines and dashed error envelopes show the maximum source-receiver offset where the impact induced seismic signal is above the noise for a given impact energy E . Grey points with errors show the number of impacts over the whole of Mars for each decadal energy bin from Table 2. Lower panels show the resulting number of detectable events per year in each energy bin at a single seismic station.