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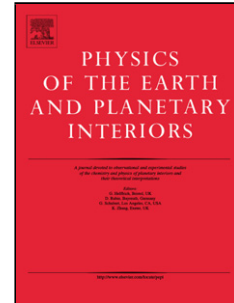
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- We model seismic signals generated by meteorite impacts on Mars.
- We predict the total seismic moment released by impacts is $1e13$ - $1e14$ Nm per year.
- Nominally, about one globally detectable impact event should occur every ten years.
- Meteorite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

Accepted Manuscript

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 **1. Introduction**

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

11 Mars has no internal magnetic field (Acuna et al., 1999; Connerney et al.,
12 1999) or reliable seismic data so currently the only way to estimate internal
13 structure is to develop models that are consistent with the measured mo-
14 ment of inertia of $I=0.365MR^2$ (Yoder et al., 2003; Sohl et al., 2005) and an
15 assumed composition based on SNC meteorites, which are believed to be rep-
16 resentative of the bulk Martian composition (Sohl and Spohn, 1997). Many

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

22 The best way to improve our knowledge of Mars' interior is with a global
23 seismic network. However, off-Earth seismology is extremely challenging.
24 The seismometer must survive the rigours of launch, landing, and harsh
25 environments with large temperature swings. Another challenge is remote
26 deployment onto the planet surface in a way that provides good seismic
27 coupling. The Apollo seismic experiment, deployed by astronauts, detected
28 numerous seismic events on the Moon, and has revolutionised our under-
29 standing of the lunar interior. Internal models now exist for lunar density
30 (Bills and Ferrari, 1977), velocity structure (Goins et al., 1981; Nakamura,
31 1983; Khan et al., 2000; Lognonne et al., 2003; Gagnepain-Beyneix et al.,
32 2006; Khan et al., 2007), and seismic attenuation (Nakamura, 1976; Goins
33 et al., 1981; Nakamura and Koyama, 1982).

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

42 Possible sources of seismicity on Mars include: (1) Fracturing caused by
43 thermal and lithostatic stresses predicted from visible faults in high resolu-
44 tion images and topographic data (Golombek et al., 1992; Knapmeyer et al.,
45 2006). These authors predict total yearly moment releases of 10^{18} Nm and
46 3×10^{16} – 5×10^{18} Nm respectively, occurring mostly as small sub-magnitude 5
47 events; (2) Volcanic events, for which there is evidence of recent (<50 Ma)
48 activity from crater counting. Sub-surface events are most likely as no cur-
49 rent surface activity is observable; and (3) Meteorite impacts, which based
50 on extrapolation of Apollo lunar data should be detectable (Davis, 1993).
51 On Earth, the majority of seismic events are due to plate tectonics - a pro-
52 cess that appears to be long since dormant on Mars - but which may have
53 operated in Mars' early history prior to 3.5Ga (Sleep, 1994).

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

61 Both ESA and NASA are considering deployment of seismic instrumen-
62 tation on Mars in the near future. Mission scenarios have included both
63 broad-band and short-period instruments (Lognonne et al., 1996, 2000; Pike
64 et al., 2005). Although Mars is predicted to be seismically active, this has
65 not yet been confirmed with seismic instruments. Therefore, the initial goal
66 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 $1\text{-}\sigma$ 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 led 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:

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

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

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

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

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

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

316 Table 2 summarises our results for the impact induced seismicity of Mars
317 and gives the number of impact events with a given D and M in each decadal
318 energy bin, with the impact rates determined from the modified Hartmann
319 (2005) isochrons. The seismic moment magnitude M_w , defined by $M_w =$
320 $2/3(\log_{10} M - 9.1)$ (Kanamori, 1977), is also given for ease of comparison
321 with terrestrial events.

322 3. Mars Interior Model

323 Current knowledge of Mars' interior is reviewed by Lognonne (2005). Here
324 we are concerned with velocity, density, and seismic attenuation as these
325 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).

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

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

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

371 5. Synthetic Seismograms

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

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

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

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

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

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

383 Modelling shallow events is too computationally expensive (Kawai et al.,
 384 2006), but is only required for surface wave generation. In this study we
 385 are only interested in body waves amplitudes, as these waves sound the deep

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

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

399 Figure 8 shows waveforms calculated for each maximum modelled fre-
400 quency for three source-receiver offsets. If waveforms are not modelled with
401 a high enough maximum frequency then “ringing” effects will be seen around
402 the arrivals and the amplitudes of the waveforms will not have converged.
403 Comparison of synthetics with different maximum frequencies was used to
404 check for amplitude convergence. Due to seismic attenuation, it was not
405 necessary to model frequencies higher than 8 Hz to accurately model syn-
406 thetic seismograms with offsets over 30° . However, closer seismograms travel
407 shorter distances and have been attenuated less. Therefore, at close range
408 higher maximum frequency is required. These effects are just visible in the
409 20° offset synthetics in Figure 8, which have not quite fully converged. There-
410 fore, 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 IGP 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-

460 tions, whereas the highest noise case represents an exploration-class instru-
 461 ment in a non-ideal deployment. These noise spectral densities n_{rms} were
 462 converted into maximum peak acceleration a_{peak} using the formula $a_{peak} =$
 463 $1.25n_{rms}\sqrt{(f_2 - f_1)}$, where f_1 - f_2 is the frequency range under consideration
 464 (Havskov and Alguacil, 2004).

465 Under nominal noise conditions ($10^{-9} \text{ ms}^{-2}\text{Hz}^{-1/2}$) Figure 11 shows that
 466 to detect an event at teleseismic distances (offset $\Delta \geq 60^\circ$) requires a seismic
 467 moment over 10^{13} Nm ($M_w=2.6$), or equivalently an impact producing a
 468 100 m diameter or larger crater. From Table 2 such large craters are produced
 469 every 1–8 years or 4 years on average. A globally detectable event would
 470 require a moment of around 10^{14} Nm ($M_w=3.3$).

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

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

474 where i_e is the number of impacts per year over the whole of Mars in each
 475 energy bin, θ is the maximum offset at which an event is detectable, and the
 476 $(1 - \cos \theta)/2$ factor gives the fractional surface area of Mars over which an
 477 event would be detectable. The nominal noise case represents the most likely
 478 scenario and shows that a single seismic station would detect impacts with
 479 $\Delta \geq 60^\circ$ approximately once every 10 years. For the low noise case, which
 480 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

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

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Quantity	Value	Fractional error	Contribution to fractional error σ_M/M
$D = aE^b$			
a	8.8×10^{-3}	$\frac{\sigma_a}{a} = 0.35$	$\frac{1}{bd} \frac{\sigma_a}{a} = 0.9$
b	0.32	$\frac{\sigma_b}{b} = 0.03$	$\frac{1}{b^2 d} \ln \left[\frac{D}{a} \left(\frac{g}{g_{\oplus}} \right)^{\frac{3}{16}} \right] \frac{\sigma_b}{b} = 2.0\dagger$
$E_s = kE$			
k	2×10^{-5}	$\frac{\sigma_k}{k} = 10$	$\frac{1}{d} \frac{\sigma_k}{k} = 8.0$
$E_s = cM^d$			
c	4.8×10^{-9}	$\frac{\sigma_c}{c} = 0.49$	$\frac{1}{d} \frac{\sigma_c}{c} = 0.4$
d	1.24	$\frac{\sigma_d}{d} = 0.01$	$\frac{1}{d^2} \ln \left[\left(\frac{D}{a} \left(\frac{g}{g_{\oplus}} \right)^{\frac{3}{16}} \right)^{\frac{1}{b}} \frac{k}{c} \right] \frac{\sigma_d}{d} = 0.2\dagger$
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}	E_{\max}	M	M_w	Diameter	N	N_{\min}	N_{\max}
(J)	(J)	(Nm)	(Nm)	(m)	(per year)	(per year)	(per year)
10^8	10^9	2.3×10^9	0.2	3.67	482	300	1240
10^9	10^{10}	1.5×10^{10}	0.7	7.64	121	65.5	316
10^{10}	10^{11}	9.3×10^{10}	1.2	15.8	19.1	9.15	70.6
10^{11}	10^{12}	6.0×10^{11}	1.8	33.0	2.30	1.09	9.98
10^{12}	10^{13}	3.8×10^{12}	2.3	68.6	0.280	0.130	1.19
10^{13}	10^{14}	2.4×10^{13}	2.9	143	0.0321	0.0140	0.142
10^{14}	10^{15}	1.6×10^{14}	3.4	297	0.00298	0.00118	0.0153
10^{15}	10^{16}	1.0×10^{15}	3.9	617	0.000220	0.000085	0.00130
10^{16}	10^{17}	6.4×10^{15}	4.5	1280	0.000018	0.000010	0.000094
10^{17}	10^{18}	4.1×10^{16}	5.0	2660	0.0000042	0.0000027	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.

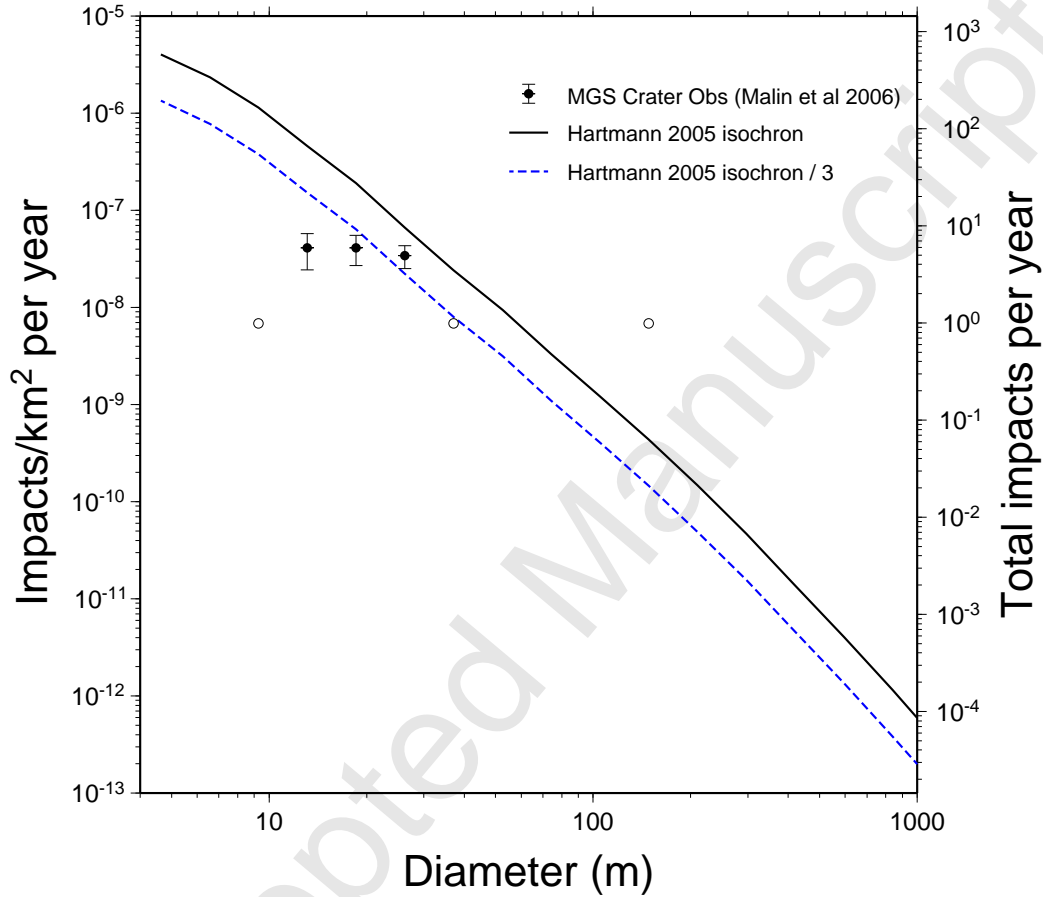


Figure 1: Cratering 1-year isochron for Mars extrapolated from the Hartmann (2005) 3 Gyr isochron compared to measurements of new craters observed by Malin et al. (2006). The observed recent cratering suggests a correction factor of 1/3 must be applied to the Hartmann (2005) isochron for recent small 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.

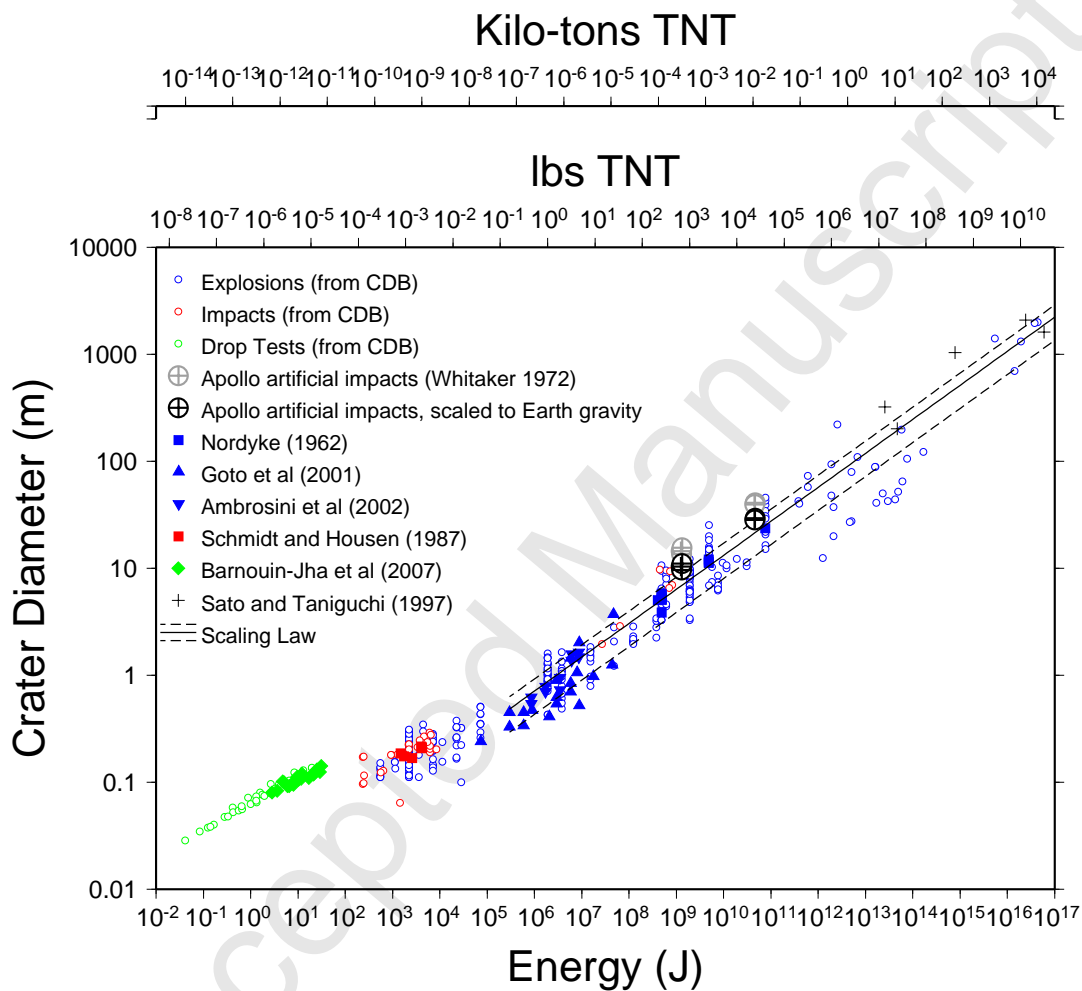


Figure 2: Measurements of crater diameter as a function of impact energy taken from the cratering database (CDB) of Holsapple et al. (2003) augmented with additional and post-1998 measurements. The solid and dashed lines show our derived scaling law and uncertainty bounds. Low energy ($E < 10^5$ J) impacts/explosions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

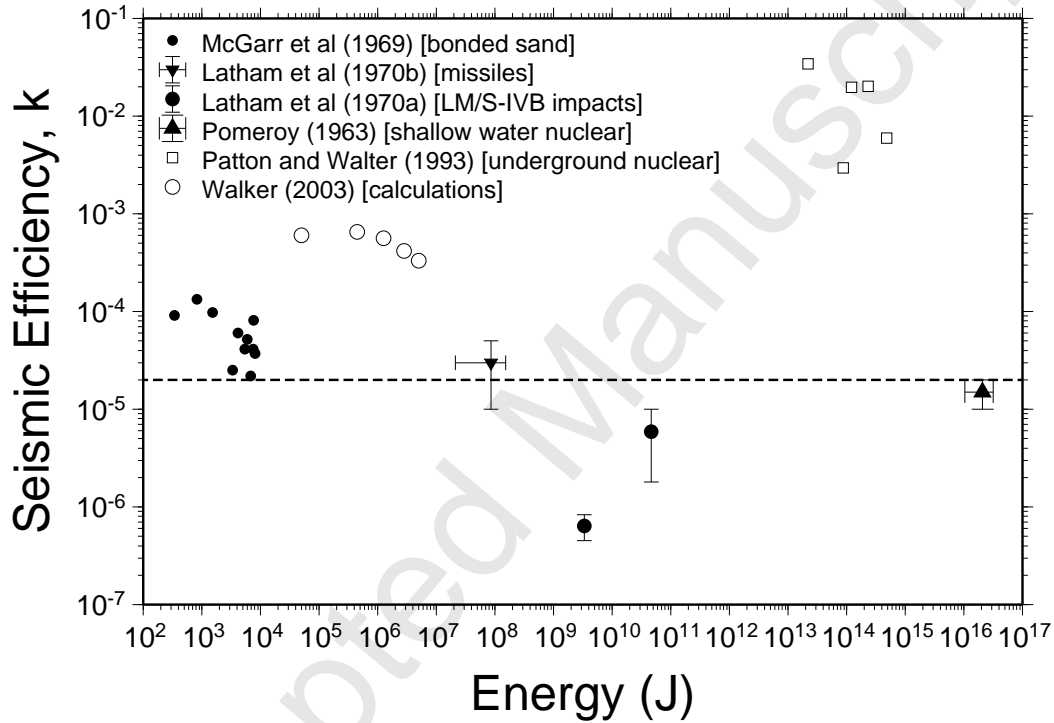


Figure 3: Seismic efficiency as a function of source energy compiled from various studies. The data are very scattered, indicating the large uncertainty in k . The dashed line shows our adopted value of $k=2 \times 10^{-5}$, which is broadly consistent (within an order of magnitude) with most of the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

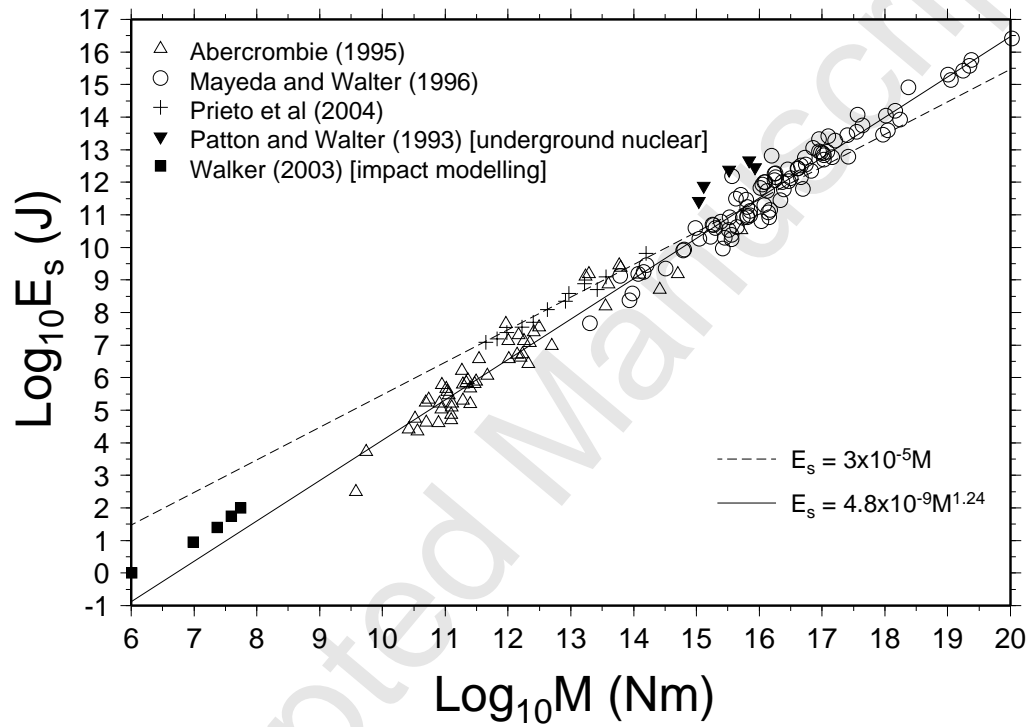


Figure 4: Relationship between seismic moment and total radiated seismic energy. Data are from regional studies of small earthquakes (Abercrombie, 1995; Mayeda and Walter, 1996; Prieto et al., 2004), impact modelling (Walker, 2003), and nuclear explosions (Patton and Walter, 1993). The trend lines are 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.

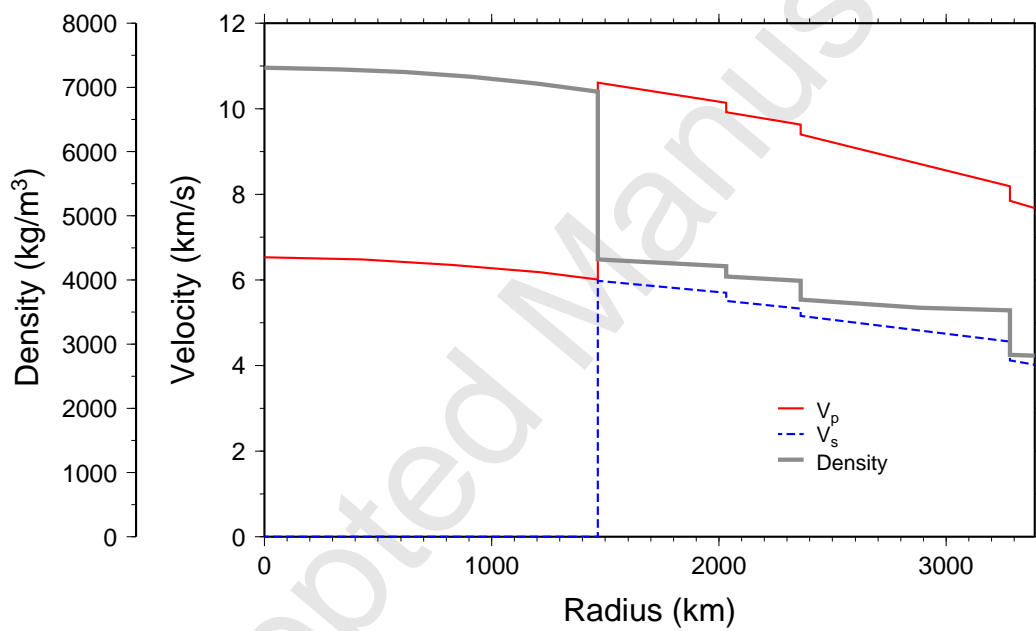


Figure 5: Seismic velocity and density model for Mars from Sohl and Spohn (1997) (Model A).

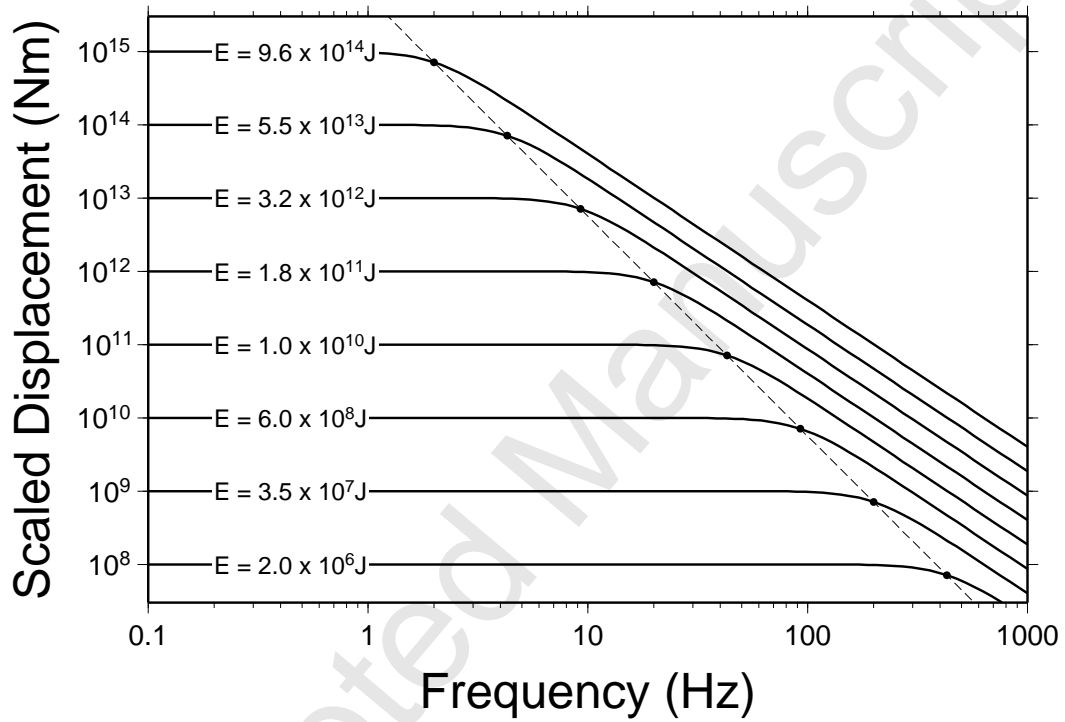


Figure 6: Seismic displacement source spectra used for creation of the synthetic seismograms scaled by the seismic moment. The equivalent impact energy E is given for each case. Source spectra are essentially flat below 4 Hz for all but largest events. Hence, at teleseismic distances the high frequency 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.

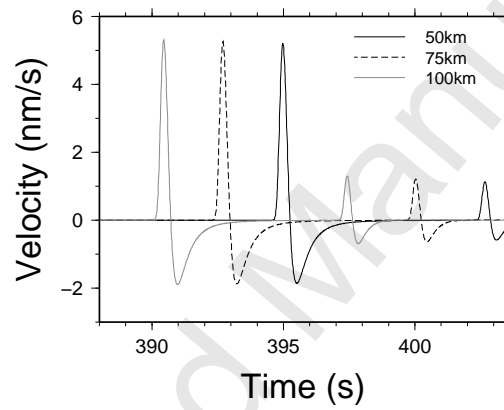


Figure 7: Synthetic seismograms for a $M=10^{14}$ Nm event generated with 8 Hz maximum frequency and source depths of 50, 75, and 100km for a source-receiver offset of 60° . The difference between body wave amplitudes (P-waves in this plot) for different source depths is negligible and can be accounted 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.

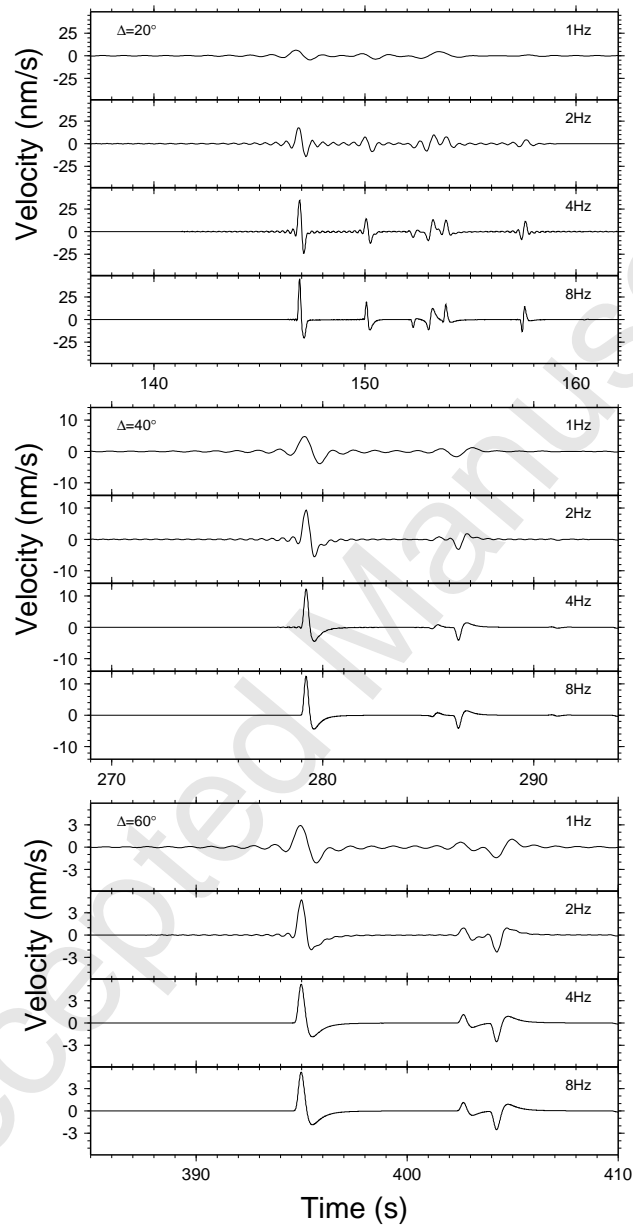


Figure 8: Waveforms for 1, 2, 4, and 8 Hz maximum modelled frequencies at source-receiver offsets Δ of 20, 40 and 60° for an event with $M=10^{14}$ Nm. Note that more distant events have lower frequency content due to increased seismic attenuation. At $\Delta=20^\circ$ some slight ringing is still visible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

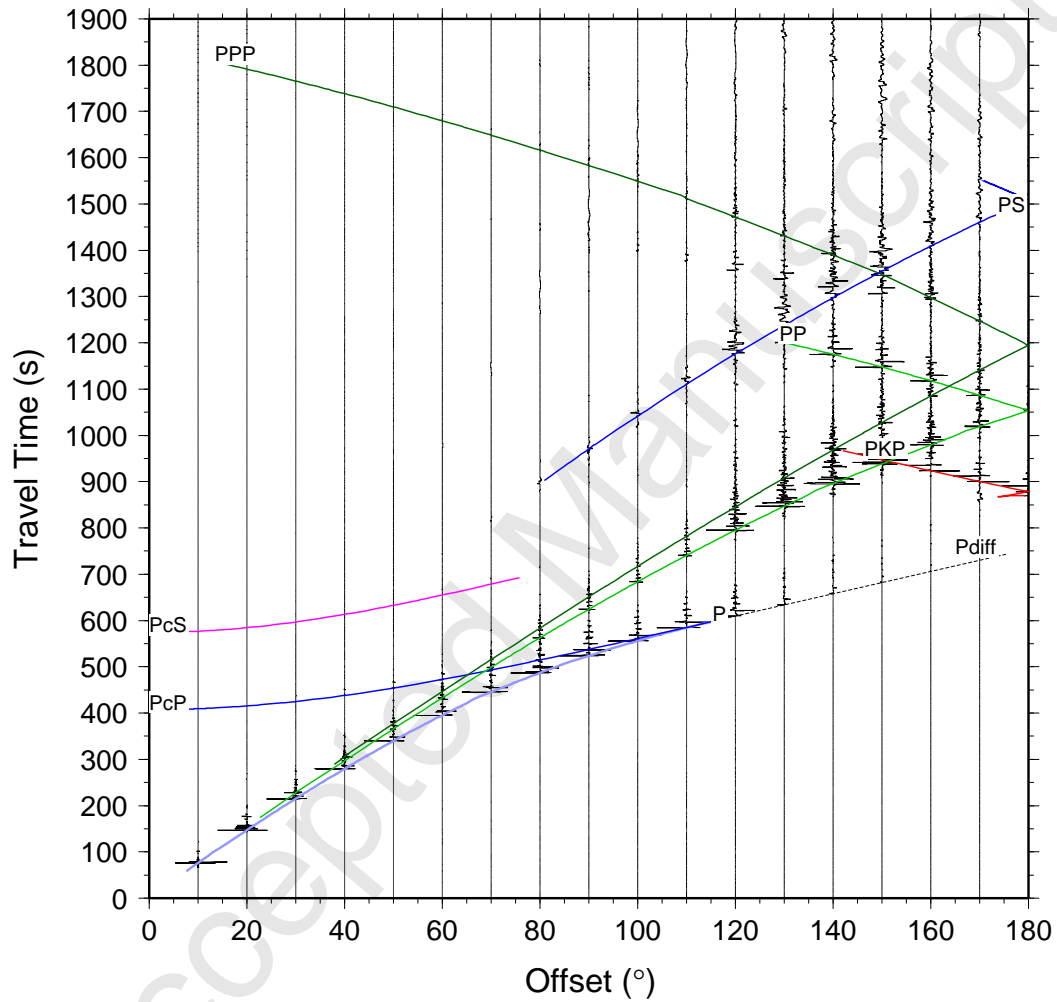


Figure 9: Record section compiled from synthetic seismograms over a full Martian hemisphere. Travel time curves calculated using the Tau-P toolkit (Crotwell et al., 1999) for various phases generated using an isotropic source are also shown. A P-wave shadow zone exists beyond $\Delta=110^\circ$.

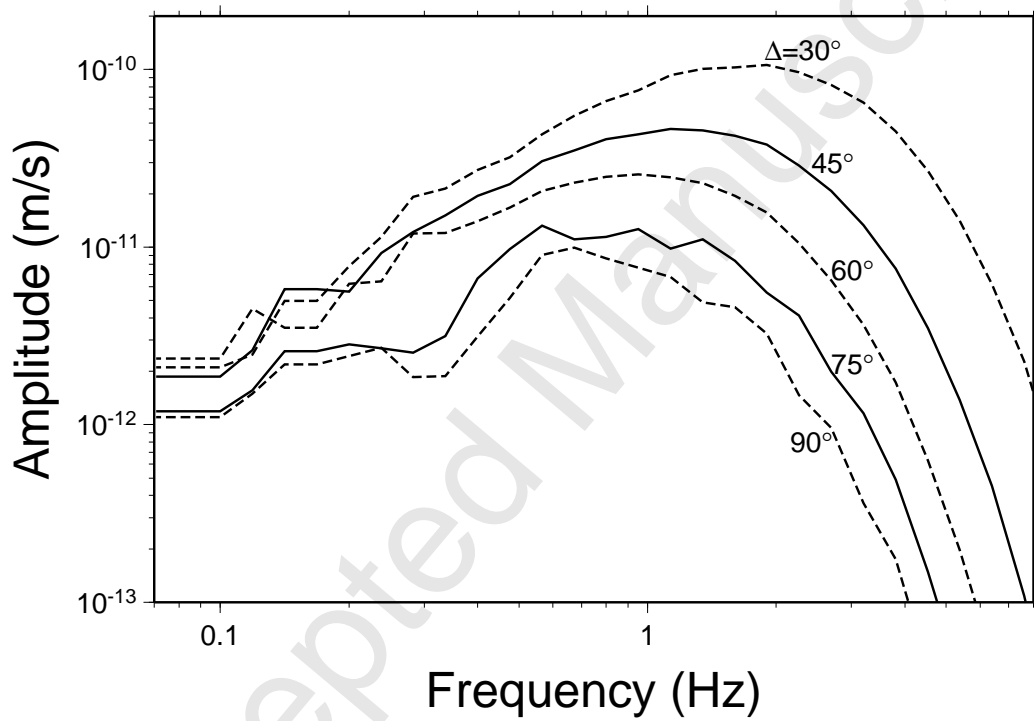


Figure 10: Spectra of P-wave envelope for a range of source-receiver offsets Δ for an event with $M=10^{14}$ Nm. Larger offsets have less high frequency content due to seismic attenuation. Most of the frequency content is in the 0.4–4 Hz range.

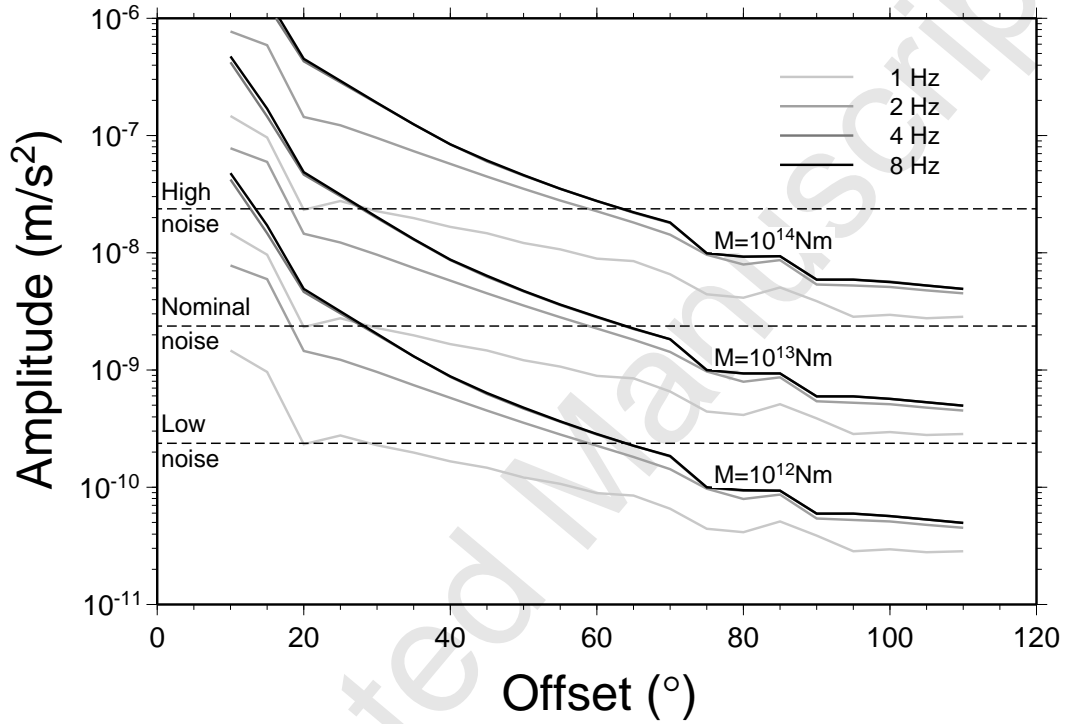


Figure 11: Maximum P-wave amplitude in the frequency range 0.4-4 Hz as a function of source-receiver offset and seismic moment for 8 Hz synthetics. Results from 1, 2, and 4 Hz models are also shown to give an idea of the convergence. Representative noise levels are also shown (dashed lines) over the 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.

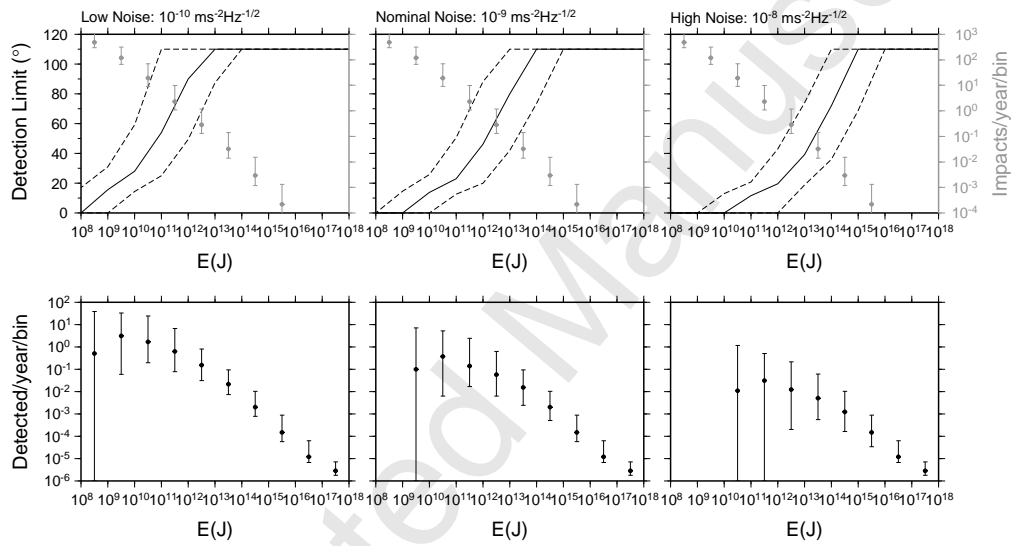


Figure 12: Detectability of meteorite impacts on Mars for deployments with low, nominal, and high noise levels. In the top panels, solid lines and dashed error envelopes show the maximum source-receiver offset where the impact induced seismic signal is above the noise for a given impact energy E . Grey 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.