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Predicted detection rates of regional-scale meteorite impacts on Mars with the InSight short-period seismometer

N. A. Teanby^a

^a*School of Earth Sciences, University of Bristol, Wills Memorial Building, Queen's Road, Bristol, BS8 1RJ, U.K.*

Abstract

In 2016 NASA will launch the InSight discovery-class mission, which aims to study the detailed internal structure of Mars for the first time. Short- and long-period seismometers form a major component of InSight's payload and have the potential to detect seismic waves generated by meteorite impacts. Large globally detectable impact events producing craters with diameters of ~ 100 m have been investigated previously and are likely to be rare (Teanby and Wookey, 2011), but smaller impacts producing craters in the 0.5–20 m range are more numerous and potentially occur sufficiently often to be detectable on regional scales ($\lesssim 1000$ km). At these distances, seismic waves will have significant high frequency content and will be suited to detection with InSight's short-period seismometer SEIS-SP. In this paper I estimate the current martian crater production function from observations of new craters (Malin et al., 2006; Daubar et al., 2013), model results (Williams et al., 2014), and standard isochrons (Hartmann, 2005). These impact rates are combined with an empirical relation between impact energy, source-receiver

Email address: `n.teanby@bristol.ac.uk` (N. A. Teanby)

distance, and peak seismogram amplitude, derived from a compilation of seismic recordings of terrestrial and lunar impacts, chemical explosions, and nuclear tests. The resulting peak seismogram amplitude scaling law contains significant uncertainty, but can be used to predict impact detection rates. I estimate that for a short-period instrument, with a noise spectral density of $10^{-8} \text{ ms}^{-2}\text{Hz}^{-1/2}$ in the 1–16 Hz frequency band, approximately 0.1–30 regional impacts per year should be detectable with a nominal value of 1–3 impacts per year. Therefore, small regional impacts are likely to be a viable source of seismic energy for probing Mars’ crustal and upper mantle structure. This is particularly appealing as such impacts should be easily located with orbital imagery, increasing their scientific value compared to other types of events with unknown origins. Finally, comparison of the empirical results presented here with the modelling study of Teanby and Wookey (2011) provides constraints on the seismic efficiency, suggesting that values of $\sim 5 \times 10^{-4}$ may be appropriate for impact generated seismic waves. Comparing explosion and impact datasets indicate that buried explosions are ~ 10 times more efficient at generating seismic waves than impacts.

Keywords: Mars, Seismology, Impacts, InSight, craters

1. Introduction

Planetary interiors have the potential to tell us a great deal about planet formation and evolution, but remain one of the great unknown frontiers in Solar System research. Currently Mars’ deep internal structure is constrained by observations of moment of inertia (Yoder et al., 2003; Sohl et al., 2005), composition estimates based on martian meteorites (Sohl and Spohn, 1997;

7 Zharkov and Gudkova, 2005), tidal dissipation inferred from the secular accel-
8 eration of Phobos (Zharkov and Gudkova, 1997), and inferences based on the
9 absence of a large-scale global magnetic field (Acuna et al., 1999; Connerney
10 et al., 1999). These observations do not uniquely constrain the internal struc-
11 ture and large uncertainties remain in fundamental properties such as core
12 size and composition. Closer to the surface, Mars’ relative crustal thickness
13 is constrained by topography and gravity data (Zuber, 2001), but this also
14 contains large uncertainties and relies on assumptions about crust-mantle
15 density contrasts.

16 The most effective way to probe a planet’s internal structure is using seis-
17 mology (Shearer, 2009), which is challenging for space missions as it requires
18 surface deployments (Anderson et al., 1976; Lognonne et al., 2000; Lorenz,
19 2012). Because of this, only Earth and the Moon currently have any reliable
20 seismic data. The Viking 2 seismometer did successfully return data, but
21 only one potential event was identified, which could have been caused by
22 wind noise due to the instrument’s unfavourable positioning on the lander
23 deck. Seismology on Mars will extend our knowledge to an intermediate sized
24 planet. This motivated NASA’s Interior Exploration using Seismic Investi-
25 gations, Geodesy and Heat Transport mission (InSight), which aims to probe
26 the detailed internal structure of Mars. InSight is due to launch in March
27 2016 and will land on Mars in September 2016, with a nominal mission length
28 of one Mars year.

29 The InSight seismometer package, SEIS, comprises two separate three-
30 axis seismometers: a short-period seismometer SEIS-SP (Pike et al., 2005;
31 Delahanty and Pike, 2014) designed to investigate frequencies above 0.1 Hz;

32 and a very broad band seismometer SEIS-VBB (Lognonne et al., 2014; Dan-
33 donneau et al., 2013) designed to investigate frequencies below 1 Hz. SEIS-SP
34 has a sample rate of 100 Hz and SEIS-VBB has a sample rate of 20 Hz. Both
35 seismometers will be mounted on a tripod that will be transferred to the
36 surface with a robot arm and protected from wind and extreme temperature
37 variations with a wind and thermal shield.

38 The two complementary seismometers will be suited to studying differ-
39 ent types of seismic event, distinguished primarily by the frequency content
40 of incoming seismic waves. Frequency content of seismic signals is depen-
41 dent upon the source mechanism, with larger events having a lower source
42 spectrum corner frequency (Shearer, 2009). There is also a dependence on
43 source-receiver distance, as higher seismic frequencies are preferentially at-
44 tenuated during wave propagation, meaning that much of the high frequency
45 content is removed from distant events. SEIS-SP has peak sensitivity to high
46 frequencies so will be most sensitive to local and regional seismic events,
47 whereas SEIS-VBB has peak sensitivity to low seismic frequencies so will be
48 most sensitive to teleseismic global events.

49 InSight’s seismometers will rely on Mars being seismically active to probe
50 the crustal and deep internal structure, so it is important to understand the
51 kinds of sources that are likely to be active and the level of that activity.
52 The two most important sources are expected to be faulting due to release
53 of crustal stress and meteorite impacts.

54 Faulting is expected to be the most significant of these sources. Com-
55 pelling evidence that faulting is still active today is provided by the fresh
56 boulder trails observed on large graben structures, which are interpreted as

57 being caused by boulders released by seismic ground shaking (Roberts et al.,
58 2012). Active faulting is also predicted from models that distribute stress
59 release from long-term cooling over the global fault population (Knapmeyer
60 et al., 2006), extrapolations from observed fault slips in units of different
61 geological ages (Golombek et al., 1992), and most recently from graben slip
62 rates determined from crater counting and high resolution topographic mod-
63 els (Taylor et al., 2013). However, the rate of seismicity is extremely uncer-
64 tain as it depends critically on the strength of the martian crust, which is
65 not well constrained. As a result, the estimated number of seismic events of
66 a given magnitude occurring per year spans five orders of magnitude (Taylor
67 et al., 2013). A fault source is also relatively complex and will be challenging
68 to fully characterise with a single seismic station (Panning et al., 2015) - the
69 depth, strike, and dip of the fault will all be unknown.

70 Meteorite impacts generate seismic energy during crater formation and
71 provide a second type of seismic source. Impacts have the advantage that
72 the seismic source is relatively simple, with an isotropic source function and
73 a surface location. It may also be possible to locate impacts using orbital
74 imagery, providing additional constraints. New impact craters have indeed
75 been observed in high resolution orbital images from both Mars Global Sur-
76 veyor (MGS) (Malin et al., 2006) and Mars Reconnaissance Orbiter (MRO)
77 (Daubar et al., 2013, 2015). Impacts were also a significant seismic source on
78 the Moon (Oberst and Nakamura, 1987; Gudkova et al., 2011), where many
79 small impacts were detected because of the low seismic noise and lack of at-
80 mosphere to ablate and decelerate incoming material. On Earth, meteorite
81 impacts are not a significant seismic source as most are ablated by the thick

82 atmosphere or obscured by high ambient noise levels. Direct seismic waves
 83 have only been detected from one natural impact event so far (Brown et al.,
 84 2008; Le Pichon et al., 2008; Tancredi et al., 2009), although seismic record-
 85 ings of airbursts are more common (Edwards et al., 2008), either as a direct
 86 airwave or a ground-coupled airwave. Mars’ thin atmosphere means that
 87 all but the smallest impactors should reach the surface, although they will
 88 be affected by ablation, deceleration, and fragmentation processes (Popova
 89 et al., 2003; Williams et al., 2014). These processes cause a deficiency in the
 90 number of small craters ($\ll 5$ m diameter) compared to airless bodies like the
 91 Moon. However, such small craters will be very difficult to detect from orbit
 92 and are unlikely to generate detectable seismic waves over any significant
 93 distance.

94 Previous studies have focused on large globally detectable impact events
 95 (Davis, 1993; Teanby and Wookey, 2011). Davis (1993) uses a scaling of
 96 the lunar results and concluded that around 20 events should be globally
 97 detectable per year. However, Hartmann (2005) showed the current martian
 98 impact rate is in fact a lot lower than assumed by Davis (1993). More re-
 99 cently, Teanby and Wookey (2011) used updated estimates of the impactor
 100 population (Hartmann, 2005; Malin et al., 2006) and seismic waveform mod-
 101 elling to predict only one globally detectable event every 10 years. However,
 102 the results were strongly dependent on the seismic efficiency, which is very
 103 poorly constrained (Schultz and Gault, 1975; Richardson et al., 2005; Teanby
 104 and Wookey, 2011) and introduces uncertainties of at least an order of mag-
 105 nitude. In any case, large, globally detectable, impact events are likely to be
 106 rare, with a rate of about one event per year if we are optimistic about the

107 seismic efficiency. These distant large impacts will have undergone signifi-
108 cant high frequency attenuation and will be best studied using SEIS-VBB
109 (Lognonne et al., 2014).

110 The paucity of large global impact events motivates the work presented
111 here. Small impacts are much more numerous (Hartmann, 2005; Malin et al.,
112 2006; Daubar et al., 2013), but provide a much weaker seismic source, mean-
113 ing that they may only be detectable regionally, which will require the seis-
114 mometer to be located in close proximity to the impact site. These events
115 will retain much of their high frequencies and be well suited to an investi-
116 gation using SEIS-SP. At such close source-receiver distances, these events
117 will be of limited use for studying the martian deep interior and core size.
118 However, they will be extremely useful for studying crustal and upper mantle
119 structure on regional scales.

120 Here I consider the detectability of small impacts based on estimates of
121 the current crater production function and the InSight SEIS-SP seismometer
122 specification. The approach is necessarily different to that in Teanby and
123 Wookey (2011); full waveform modelling of local/regional small impacts is
124 computationally unfeasible as it would require modelling of high seismic fre-
125 quencies. Such modelling would also be dependent on crustal model assump-
126 tions, which contain large uncertainties and are likely to be highly variable.
127 Instead, analogue terrestrial and lunar data from impacts and explosions are
128 used to empirically determine signal levels and associated uncertainties for a
129 given impact event.

130 Section 2 compiles updated estimates of the current crater production
131 function based on recent new crater detections. Section 3 determines the

132 relation between impact energy, source-receiver distance, and seismic ground
133 velocity for a wide range of crustal settings. Section 4 then develops a scaling
134 relation between crater size and detection range, which is used to predict the
135 number of events that could be detected with InSight’s SEIS-SP instrument.
136 Implications and limitations of the analyses are considered in Section 5

137 **2. Current Crater Production Function**

138 The current crater production rate on Mars can be defined in terms of
139 isochrons. Throughout this study I follow Hartmann (2005)’s definition of
140 an isochron, which is the incremental number of new craters in a given di-
141 ameter range (or bin) created in a specified time interval. Conventionally,
142 the bin centres are spaced equally in logarithmic space by a factor of $\sqrt{2}$,
143 so that a bin centred on crater diameter D includes craters with diameters
144 from $2^{-1/4}D$ to $2^{1/4}D$. Hartmann (2005) determined the crater production
145 isochrons for Mars over geological timescales based on extrapolations from
146 dated lunar samples. Subsequently, high resolution orbital imaging cam-
147 paigns have discovered many new impact sites (Malin et al., 2006; Daubar
148 et al., 2013, 2015), which provide an independent measure of the current
149 cratering rate. New impact sites are typically first identified as low-albedo
150 impact streaks in dusty areas, interpreted as clearing of higher albedo sur-
151 face dust by the impact blast. These dark streaks are much easier to identify
152 than craters alone, as they are much larger and are easily identified in sin-
153 gle images. Once a potential new impact site is identified, repeat images (if
154 available) can be used to check if the crater is indeed new. For this reason,
155 most new impact site discoveries have been restricted to dusty regions.

156 Malin et al. (2006) used wide angle Mars Orbiter Camera (MOC) images
 157 from MGS with a resolution of 230 m/pixel to search for dark spots from
 158 new impacts. Follow up images were taken with the narrow angle camera at
 159 a resolution of 1.5 m/pixel. This allowed 20 new impact sites to be identified,
 160 which had crater diameters from 2–148 m. However, the largest 148 m crater
 161 is now suspected to be much older as aeolian bedforms are visible in crater
 162 bottom (Daubar et al., 2013). More recently Daubar et al. (2013) used MRO
 163 Context Camera (CTX) images with resolutions of 6 m/pixel combined with
 164 images from Viking, Mars Odyssey, Mars Express, MGS, and MRO to search
 165 for potential new impact sites. From these, 248 impacts sites were confirmed
 166 as new following inspection of 0.25 m/pixel High Resolution Imaging Science
 167 Experiment (HiRISE) images from MRO. A subset of 44 sites were particu-
 168 larly well constrained as they had both before and after CTX images. Crater
 169 diameters of 1.7–34 m were measured for this subset using HiRISE images
 170 (Daubar et al., 2013, their Table 1).

171 Note that fragmentation in Mars’ atmosphere often causes clusters of
 172 impact craters at a given impact site (e.g. Popova et al., 2003). The diameters
 173 reported by both Malin et al. (2006) and Daubar et al. (2013) are in terms of
 174 the so-called “effective diameter” D_{eff} , which represents an equivalent single
 175 crater diameter that would be created if no fragmentation occurred. D_{eff} is
 176 defined by:

$$D_{eff} = \left(\sum_{i=1}^n D_i^3 \right)^{1/3} \quad (1)$$

177 for a cluster with n individual craters with diameters D_i . Throughout the
 178 rest of this study I treat clusters of craters as a single crater with the effective

179 diameter. However, Daubar et al. (2013) report that 56% of new impact sites
180 comprise crater clusters. The effect of this on potential impact-generated
181 seismic signals is considered further in Section 5.

182 The new crater observations studies show that smaller craters are gener-
183 ally more numerous, as expected from the isochrons and crater populations.
184 However, at the very smallest diameters, both Malin et al. (2006) and Daubar
185 et al. (2013) observe a reduction in the number of small craters, which begins
186 at ~ 10 m in the Malin et al. (2006) study and ~ 5 m in the Daubar et al.
187 (2013) study. This downturn has two possible origins: atmospheric ablation
188 removing the smallest impactors (Popova et al., 2003; Williams et al., 2014);
189 or detection biases due to finite image resolution, which makes smaller craters
190 more difficult to detect. Popova et al. (2003) predict that atmospheric effects
191 should cause a downturn in small craters beginning at around 5 m, but more
192 recent modelling by Williams et al. (2014) place the downturn at a crater
193 diameter of around 0.2 m. As the CTX image resolution used to detect the
194 new impacts is 6 m/pixel it is not possible to tell if the downturn at small
195 diameters observed by Daubar et al. (2013) is due to atmospheric effects or
196 finite image resolution. However, very high resolution HiRISE crater counts
197 at Zunil crater by Williams et al. (2014) have a downturn starting around
198 1–2 m, comparable to the minimum crater size detectable with HiRISE, sug-
199 gesting that atmospheric effects do not significantly reduce small cratering
200 events until the sub-metre scale or below.

201 The new craters observations are now used to estimate present-day crater
202 production functions. First, the new craters reported in Daubar et al. (2013)
203 (their Table 1) and Malin et al. (2006) (their Table S1) are binned into

204 the same $\sqrt{2}$ crater diameter bins as Hartmann (2005). Second, crater
 205 numbers were rescaled by the area-time function (ATF) defined in Daubar
 206 et al. (2013) to give the production function in units of impacts/km²/yr
 207 (ATF=143 499 219 km²yr for Malin et al. (2006), ATF = 19 718 204 km²yr
 208 for Daubar et al. (2013)). Third, to determine cratering rates for larger di-
 209 ameter craters, whose formation has so far not been observed, the Hartmann
 210 (2005) 1 Gyr isochron was extrapolated to a 1 yr isochron by multiplying
 211 the incremental crater numbers by 10^{-9} , followed by a further rescaling by
 212 1/3 in order to match the observational data. This rescaling is the same as
 213 used by Teanby and Wookey (2011), which only used the Malin et al. (2006)
 214 observations.

215 Because of the uncertainty surrounding the diameter threshold for atmo-
 216 spheric suppression of small crater diameters, I consider two impact models
 217 for the present-day crater production function:

218 **Impact model 1:** This model represents a lower bound on the present-
 219 day crater production function by assuming that the Daubar et al. (2013)
 220 observations represent the full extent of the cratering process. For craters
 221 falling in the 13.08 m diameter bin or above, the cratering rate is assumed
 222 to be given by the rescaled Hartmann (2005) 1 yr isochron. In addition to
 223 allowing extrapolation to larger crater diameters, using the rescaled isochron
 224 reduces scatter caused by small crater counts in the larger diameter bins.
 225 Error-bars are assumed to be the standard factor of 2 error discussed in
 226 Hartmann (1999, 2005). For craters in the 9.29 m bin and below the Daubar
 227 et al. (2013) results have sufficient counting statistics to be used directly.
 228 Error-bars are calculated from the Poisson statistics of the counts. This

229 model assumes that all the downturn at small crater diameters is caused by
230 atmospheric ablation at a level consistent with the predictions by Popova
231 et al. (2003) and has been fully resolved by the Daubar et al. (2013) new
232 crater detections.

233 **Impact model 2:** This model represents an attempted best guess at
234 the present-day crater production function, by assuming that atmospheric
235 ablation effects are consistent with the most recent Williams et al. (2014)
236 modelling. For craters falling in the 18.57 m diameter bin or above, the
237 cratering rate is assumed to be given by the Hartmann (2005) isochron after
238 rescaling to fit the Daubar et al. (2013) observations. For craters in the
239 13.08 m bin and below, the cratering rate is assumed to be given by the
240 Monte Carlo model ablation/deceleration/fragmentation results presented in
241 Williams et al. (2014) (their Figure 7a), after rescaling to fit the Daubar
242 et al. (2013) cratering rate observed in the 9.29 and 13.08 m bins. Error-bars
243 for the entire curve are assumed to be given by the standard factor of 2 error
244 discussed in Hartmann (1999, 2005).

245 The resulting composite crater production functions are shown in Figure 1
246 and specified in Table 1. Certain caveats apply to these impact models: (1)
247 using the observational studies implies that all new craters in the regions
248 studied were identified; (2) the production functions may have varied in the
249 past (Quantin et al., 2007; JeongAhn and Malhotra, 2014), so are only ap-
250 plicable to the present day; and (3) the impact rate may be dependent on
251 Mars' orbital phase as the high orbital eccentricity takes Mars closer or fur-
252 ther from the asteroid belt depending on the season (Daubar et al., 2012).
253 Caveat (1) is the most important for this study, whereas (2) operates on

254 multi-millennial timescales, far beyond the scope of a space mission, and (3)
 255 may result in changes that are not resolvable over the course of the mission if
 256 impact detection rates are low. Note that for craters diameters $\gtrsim 10$ m both
 257 impact models are lower by 1/3 than Hartmann (2005)'s isochrons, although
 258 they are just about consistent with the uncertainties he originally proposed.
 259 The difference could be due to uncertainties in the relative impactor source
 260 population on Mars compared to the Moon, or due to some new craters es-
 261 caping detection with CTX. However, for the purposes of determining seismic
 262 detection rates, the impact models presented here provide a reasonable and
 263 somewhat conservative estimate of current cratering rates.

264 **3. Estimation of Seismogram Amplitude**

265 When studying seismograms, we are primarily interested in detecting the
 266 first arrival as this is often the most distinct. SEIS-SP is a velocity sensor
 267 and measures the ground velocity caused by seismic waves. To estimate the
 268 peak ground velocity of the first arrival for a particular impact I use ana-
 269 logue data from terrestrial impacts, lunar impacts, and terrestrial explosions
 270 to determine an empirical relation between impact energy, source-receiver
 271 distance, and maximum seismogram amplitude. A range of continental set-
 272 tings are considered in order to obtain sufficient statistics and uncertainty
 273 estimates. The variability in these data are representative of variability in
 274 Earth's crustal properties and source coupling and so also provides a mea-
 275 sure of potential variability on Mars. Where possible I used broadband or
 276 extended short-period recordings with sample rates of 100Hz to be most
 277 directly comparable with the SEIS-SP instrument and to ensure that high

278 seismic frequencies were captured. The impact dataset is relatively small and
 279 comprises the Bolivian Carancas event, artificial lunar impacts, and missile
 280 impact tests. These data are supplemented with chemical and nuclear explo-
 281 sion data, which are often considered as a close analogue to impact sources
 282 (Teanby and Wookey, 2011). Source-receiver offsets up to 1200km are consid-
 283 ered, i.e. regional data dominated by crustal and upper mantle structure that
 284 will be largely insensitive to differences in deep internal structure (Kennett,
 285 2003).

286 *3.1. Impact Data*

287 *3.1.1. Lunar Apollo Artificial Impacts*

288 The Apollo seismic experiment is summarised in Latham et al. (1969,
 289 1970a) and Nakamura et al. (1982) and included both long- and short-period
 290 seismometers deployed by the astronauts. Many natural and artificial im-
 291 pacts were detected with the Apollo seismometers (Oberst and Nakamura,
 292 1987; Gudkova et al., 2011). Here I only consider artificial impacts by the
 293 spent Saturn V Apollo booster stage (SIVB) and the ascent stage of the lunar
 294 module (LM) as they have known impact velocities, masses, locations, and
 295 times, so provide a set of controlled sources with known properties (Table 2).

296 For closest comparison with SEIS-SP, I consider the Apollo short-period
 297 seismometer data, which was operated with a sample rate of around 48 Hz.
 298 Many of these recordings had low signal-to-noise, which make identification
 299 of the first arrivals and amplitudes difficult. Therefore, I selected the three
 300 impacts with the largest signal-to-noise, which were the Apollo 14 LM, the
 301 Apollo 16 SIVB, and the Apollo 17 SIVB. The locations of these impacts in
 302 relation to the Apollo seismometers are shown in Figure 2a. Seismic data

303 from the impacts recorded on the Apollo 14, 15, and 16 seismometers are
304 shown in Figure 3.

305 3.1.2. *Bolivian Carancas Impact Crater*

306 The Carancas impact event occurred on 15th September 2007 at 16:40:14
307 UT and produced a 13.5 m diameter crater (Brown et al., 2008; Le Pichon
308 et al., 2008; Tancredi et al., 2009). Estimates of source properties are sum-
309 marised in Table 2. Seismic waves generated by the impact were recorded on
310 the Bolivian Seismic Network (BSN) of 1 Hz short-period sensors at 50 Hz
311 sample rate and on the Global Seismic Network (GSN) LPAZ station broad-
312 band and short-period sensors at 40 Hz sample rate. I consider the LPAZ
313 data as slightly more reliable than the BSN data, as the sensor specification
314 is higher. Also, absolute amplitudes can be more reliably recovered from
315 LPAZ as full instrument transfer functions were available, whereas only ap-
316 proximate sensor gains were available for the BSN stations. The location of
317 the impact is shown in Figure 2b and seismic data for the impact are shown
318 in Figure 4. The Carancas impact event is the only example of direct seismic
319 waves from an impact event on Earth. Other events have only been recorded
320 seismically via ground coupling of an associated airburst. It is important
321 to note that the impact occurred in water saturated soil and may have pro-
322 duced a larger crater than would have been created in solid rock. Tancredi
323 et al. (2009) estimate that an impact energy of 1000–3000 kg TNT would be
324 required to form the crater in this terrain. Soil water saturation could also
325 effect the efficiency of seismic wave generation and is a sub-optimal analogue
326 for the surface of Mars.

327 3.1.3. *Nevada Missile Tests*

328 In preparation for the Apollo seismic experiment Latham et al. (1970b)
329 investigated seismic signals generated by five missile impacts at White Sands
330 Nevada test site during 1968 and 1969. Seismograms were recorded with
331 small geophones on analogue equipment and were used to determine the max-
332 imum ground displacements for P-waves and Rayleigh waves (Latham et al.,
333 1970b, their Table 2). P-wave displacements were converted into maximum
334 velocity amplitudes using a harmonic wave approximation and the reported
335 dominant frequencies. Unfortunately the raw seismic data and instrument
336 specifications are not available, so this approximation will introduce some
337 uncertainty.

338 3.2. *Explosion Data*

339 3.2.1. *EAGLE Chemical Explosions*

340 The Ethiopia-Afar Geoscientific Lithospheric Experiment (EAGLE) is a
341 large international project to study the Ethiopian segment of the east African
342 rift (Maguire et al., 2003). Part of the project involved a controlled source
343 phase in January 2003, where 23 explosive sources with yields from 50–
344 5750 kg TNT were used to image the rift (Maguire et al., 2006). I used
345 the 11 shot points with the highest signal-to-noise, which are summarised in
346 Table 2. Explosions were recorded on a dense network of ≈ 1000 geophones
347 and 93 broadband sensors covering an area approximately 300 x 300 km
348 shown in Figure 2c. Here, I only consider the broadband data, which were
349 recorded at 100 Hz on Guralp 6TD sensors. The combination of dense cov-
350 erage with broadband instruments and a large number of explosive sources
351 means this dataset is extremely well suited to studying regional amplitude

dependence.

3.2.2. Nuclear Explosions

To extend the EAGLE chemical explosion dataset to larger yields and greater source-receiver distances I also consider seismic data from nuclear tests in the US, China, and North Korea (Democratic People’s Republic of Korea, DPRK). The analysis is restricted to tests conducted after 1990, where high quality seismic data are available, and to tests with reliable source yield estimates. Source parameters for these tests are given in Table 2 and locations are shown in Figure 2d–f.

3.3. Data Extraction and Processing

All data were extracted from the Incorporated Research Institutions for Seismology (IRIS) database in full Standard for the Exchange of Earthquake Data (SEED) format, except for Bolivian Seismic Network data, which were obtained directly from the Observatorio San Calixto, Bolivia (E. Minaya *pers. comm.*) in Group of Scientific Experts (GSE) format. For ease of manipulation, SEED data were converted into Seismic Analysis Code (SAC) format (Goldstein et al., 2003; Helffrich et al., 2013) using the `rdseed` utility from IRIS. An initial visual quality control step was performed using SAC to remove very noisy, clipped, or otherwise corrupted data. Instrument responses were deconvolved using the response (RESP) files supplied with SEED volumes, during which a frequency taper was applied to limit deconvolution to frequencies within the instruments’ response range and prevent deconvolution instabilities. The exception to this was the BSN data, which were simply rescaled with the supplied linear sensor gains. Deconvolution converted raw

376 sensor counts and voltages into physical velocity units (ms^{-1}). Below 1 Hz
377 recordings were generally contaminated by microseismic noise. Therefore, a
378 1–16 Hz 4 pole Butterworth filter (Gubbins, 2004) was applied to remove
379 microseismic noise, long-period instrument drift, and high frequency noise,
380 which helped identification of the first arrivals. The peak seismogram am-
381 plitude, i.e. the maximum ground velocity of the first arrival, was obtained
382 from the various datasets as follows.

383 First, data were formed into common shot gathers and sorted in order
384 of source-receiver distance to form record section plots as in Figure 5. This
385 allowed identification of different arrival phases by comparison with standard
386 travel time curves. Here I focus on the first arrivals, which on regional scales
387 are crustal and mantle phases such as Pg and Pn (Kennett, 2003).

388 Second, a first arrival time window and a preceding noise time window
389 were defined in order to estimate the maximum amplitude and associated
390 uncertainty. Datasets were small enough that first arrival and noise windows
391 could be picked by hand for all datasets except EAGLE, where there were
392 over a thousand seismograms. For the EAGLE data, I used a window ± 5 s
393 relative to the shot origin time in terms of reduced time (Shearer, 2009)
394 assuming a p-wave velocity of 6km/s (following Maguire et al., 2006). The
395 noise window was defined as 5–20 s before the reduced origin time. This was
396 manually checked for each shot’s record section to make sure that the first
397 arrivals were included in the time window. The emergent lunar seismograms
398 did not have a distinct first arrival phase so a first arrival window length of
399 20 s was chosen.

400 Third, the peak signal amplitude in the first arrival window was extracted,

401 along with the peak noise amplitude in the noise window, which was used to
 402 assign an uncertainty to each peak amplitude. Note that for the lunar seis-
 403 mograms the choice of a 20 s first arrival window length has an effect on the
 404 peak signal amplitude, with longer windows resulting in higher peak signal
 405 (reasonable choices give consistent results to within a factor of two). Am-
 406 plitudes with a signal-to-noise ratio less than five were rejected from further
 407 analysis. I also examined the maximum of the envelope function obtained
 408 from the Hilbert transform of the seismogram (Shearer, 2009). However no
 409 significant difference between peak amplitudes and envelope function max-
 410 ima were observed in the filtered seismograms, so I used peak amplitude for
 411 simplicity.

412 Therefore, for each individual seismogram, the above procedure resulted
 413 in a source yield, a source-receiver distance, a first arrival peak ground veloc-
 414 ity, and an associated uncertainty. The results from the continental settings
 415 used in this study should be broadly transferable to Mars (see Section 5).
 416 However, care must be taken in the case of the EAGLE study, where active
 417 rifting is occurring. Maguire et al. (2006) found normal crystalline crust away
 418 from the rift’s central axis, but report strong reverberations due to possible
 419 intrusions in the rift centre, which caused anomalously high amplitudes at
 420 80–180 km offsets. These offsets were rejected from further analysis.

421 *3.4. Distance-Yield-Amplitude Relation*

422 Analysis of the seismic data resulted in a set of impact or explosion yields
 423 y_0 (kg TNT equivalent), source-receiver distances x_0 (km), maximum ground
 424 velocities v (ms^{-1}), and uncertainties σ_v (ms^{-1}). Inspection of this data

425 suggested a power law dependence of the form:

$$v(x_0, y_0) = a_0 x_0^b y_0^c \quad (2)$$

426 where a_0 , b , and c are empirically derived constants. The physical meaning
 427 of these parameters is as follows. Parameter a_0 is directly proportional to
 428 the seismic efficiency k_s , which is the fraction of the impact or explosion
 429 energy converted into seismic energy. Parameter b should have a value of
 430 approximately -1 for spherically propagating waves in an isotropic medium
 431 with no attenuation. However, attenuation will reduce the value of b and
 432 non-spherical propagation due to crustal velocity gradients will also affect
 433 b . Parameter c , the dependence on yield, has been much debated in the
 434 literature, with values in the range $1/3$ – 1 being suggested (see discussion in
 435 Kohler and Fuis, 1992). The most comprehensive study is that of Larson
 436 (1982), who found a value of around $1/3$ using yields spanning 10 orders of
 437 magnitude in sodium chloride (laboratory scales to large nuclear tests). If
 438 only energy conservation is considered, a value of $c = 1/2$ would be expected,
 439 as the kinetic energy of an elastic seismic wave is proportional to the ground
 440 velocity squared.

441 The parameters a_0 , b , and c were fitted to the explosive dataset using
 442 unweighted linear least squares (Gubbins, 2004). The fitted value of c was
 443 0.49 ± 0.03 , which is indistinguishable from the idealised energy conservation
 444 value of $1/2$. Therefore, c was fixed at $1/2$ and the other parameters were
 445 refitted to simplify subsequent analysis. The fit parameters and uncertainties
 446 are given in Table 3. Figure 6 shows the fitted relationship for $v(x_0, y_0)$,
 447 where the velocity has been rescaled to that of a standard 1000 kg TNT

448 source using:

$$v(x_0, 1000 \text{ kg TNT}) = v(x_0, y_0) \left(\frac{1000 \text{ kg TNT}}{y_0} \right)^{1/2} \quad (3)$$

449 The relationship is linear (in logarithmic space) over a wide range of source-
450 receiver distances and explosive yields. Unfortunately, there are not enough
451 impact data to reliably fit the parameters a_0 , b , and c to impacts alone.
452 Therefore, my approach is to use the explosive data as a basis for extrapola-
453 tion to the impact case.

454 Further consideration is required before translating the fitted parameters
455 from explosive sources to impacts sources. As shown in Figure 6, explosions
456 generally give higher peak velocities than impacts. This is primarily because
457 explosives are buried to maximise the seismic coupling in controlled source
458 experiments like EAGLE; or in the case of nuclear weapons testing, to avoid
459 surface damage and undesirable radioactive fallout. Therefore, explosives
460 tend to have a higher seismic efficiency and a correspondingly higher value for
461 the a_0 parameter (see also discussion of seismic efficiency in Richardson et al.,
462 2005; Teanby and Wookey, 2011). However, the effect of distance is entirely
463 dependent on crustal properties and wave propagation, so parameter b can
464 be assumed to be the same for both impacts and explosions. Note that the
465 Moon is known to have a high seismic Q (low seismic attenuation) (Nakamura
466 and Koyama, 1982; Lognonne and Mosser, 1993; Lognonne et al., 2003) and
467 emergent arrivals due to crustal scattering. Inspection of Figure 6 shows
468 that the lunar impacts have the same source-receiver distance dependence as
469 terrestrial explosions, so the high Q does not significantly affect the distance
470 dependence of peak seismogram amplitudes on regional scales. However, this
471 could be a coincidence due to competing effects of high lunar Q and high

472 crustal scattering. Parameter c should also have a similar value for impacts
 473 and explosions due to the similarities between these two source types; both
 474 sources are isotropic, effectively occur at a point source, and preferentially
 475 generate P-waves.

476 Therefore, to first order the relationship for $v(x_0, y_0)$ can be translated
 477 from explosions to impacts via application of a simple scale factor s to the a_0
 478 parameter. To allow constraints determined from the larger explosion dataset
 479 to be used b and c were fixed and the scale factor s was fitted to the impact
 480 dataset. The best fitting value of s is 0.099, implying that buried explosions
 481 are ~ 10 times more effective at generating seismic waves than impacts. The
 482 value of s contains an order of magnitude uncertainty due to the sparse and
 483 varied nature of the impact data. This large uncertainty is inevitable as
 484 s depends linearly on the impact seismic efficiency. The uncertainty in s
 485 places the upper error bound in line with the Apollo impact results and the
 486 lower bound in line with the LPAZ Carancas measurements. Parameters and
 487 uncertainties for the $v(x_0, y_0)$ relation for impacts are given in Table 3. Note
 488 that this relationship is only valid for events with source-receiver distances
 489 of < 1200 km. Also, as the raw data is bandpass filtered between 1–16 Hz,
 490 this relationship is only applicable for frequencies within the 1–16 Hz range.

491 4. Regional Impact Detection

492 Consider an impact with yield y_0 (kg TNT equivalent) a distance x_0 (km)
 493 from the SEIS-SP seismometer. From Section 3 the peak ground velocity
 494 $v(x_0, y_0)$ (ms^{-1}) of the first arrival from is:

$$v(x_0, y_0) = a_0 s x_0^b y_0^c \quad (4)$$

495 The yield of TNT is $q=4.18 \times 10^6$ J/kg (Shoemaker, 1983), so in SI units
 496 equation 4 becomes:

$$v(x, y) = a_0 s \left(\frac{1}{1000} \right)^b \left(\frac{1}{q} \right)^c x^b y^c \quad (5)$$

$$(6)$$

497 where x has units of meters and y has units of Joules. This can be simplified
 498 by setting:

$$a = a_0 s \left(\frac{1}{1000} \right)^b \left(\frac{1}{q} \right)^c \quad (7)$$

499 which gives:

$$v(x, y) = ax^b y^c \quad (8)$$

500 Constraints on the impact rate from section 2 are in terms of crater diameter
 501 D (meters), which is related to impact energy y (Joules) by (Teanby and
 502 Wookey, 2011):

$$D = \alpha_{\oplus} y^{\beta} \left(\frac{g_{\oplus}}{g} \right)^{3/16} \quad (9)$$

503 where $\alpha_{\oplus}$ and β are empirically derived constants, $g_{\oplus}$ is Earth's gravity, and
 504 g is the gravity on Mars. Setting:

$$\alpha = \alpha_{\oplus} \left(\frac{g_{\oplus}}{g} \right)^{3/16} \quad (10)$$

505 Gives $D = \alpha y^{\beta}$ and:

$$y = \left(\frac{D}{\alpha} \right)^{1/\beta} \quad (11)$$

506 So, in terms of crater diameter, the peak ground velocity on Mars is:

$$v(x, D) = ax^b \left(\frac{D}{\alpha} \right)^{c/\beta} \quad (12)$$

507 As is conventional for broadband seismometers, the noise level of SEIS-SP is
 508 specified in terms of acceleration noise power spectral density p_a ($\text{ms}^{-2}\text{Hz}^{-1/2}$).
 509 The corresponding velocity noise power spectral density p_v ($\text{ms}^{-1}\text{Hz}^{-1/2}$) at
 510 frequency f is:

$$p_v = \frac{p_a}{2\pi f} \quad (13)$$

511 The peak velocity noise n_v (in ms^{-1}) in the frequency range f_1 – f_2 is given
 512 by (Havskov and Alguacil, 2004):

$$n_v = \frac{5}{4} p_v \sqrt{f_2 - f_1} \quad (14)$$

513 Combining equations 13 and 14 with the geometric mean central frequency
 514 $f = \sqrt{f_1 f_2}$ gives:

$$n_v = \frac{5}{8\pi} p_a \sqrt{\frac{1}{f_1} - \frac{1}{f_2}} \quad (15)$$

515 For an event to be detectable the signal must be greater than the noise.
 516 Therefore, the maximum source-receiver distance x_{max} where an impact is
 517 detectable is given by the criteria:

$$v(x_{max}, D) = n_v \quad (16)$$

518 Combining equations 12, 15 and 16 gives the detection criteria:

$$ax_{max}^b \left(\frac{D}{\alpha} \right)^{c/\beta} = \frac{5}{8\pi} p_a \sqrt{\frac{1}{f_1} - \frac{1}{f_2}} \quad (17)$$

519 Therefore,

$$x_{max}(D) = \left(\frac{5}{8\pi} p_a \sqrt{\frac{1}{f_1} - \frac{1}{f_2}} \right)^{1/b} \alpha^{c/b\beta} a^{-1/b} D^{-c/b\beta} \quad (18)$$

520 Equation 18 gives the criteria for the maximum detection range $x_{max}(D)$ of
 521 an impact crater with diameter D for an instrument with an acceleration
 522 noise power spectral density of p_a in the frequency range $f_1 - f_2$. If the
 523 source-receiver distance x is less than x_{max} then the impact will be detectable,
 524 otherwise it will be below the instrument noise. The values of the parameters
 525 and their fractional errors are given in Table 3. Note that parameters a , b ,
 526 and c are potentially frequency dependent and application of equation 18
 527 outside the 1–16 Hz frequency range would require these parameters to be
 528 redetermined.

529 To determine the impact detection rate with SEIS-SP the crater produc-
 530 tion functions from section 2 must be combined with the detection criteria
 531 in equation 18. First, equation 18 is used to determine the maximum detec-
 532 tion range of a crater with a given diameter. I use the SEIS-SP instrument
 533 specification of $p_a=10^{-8} \text{ ms}^{-2}\text{Hz}^{-1/2}$ (Lognonne et al., 2014) and a frequency
 534 range of 1–16 Hz, which corresponds to that used to determine the parame-
 535 ter values and is appropriate for regional events. Fractional parameter errors
 536 are propagated through equation 18 using the formulae in Bevington and
 537 Robinson (1992) (Figure 7a). The detection range predictions are reliable up
 538 to source-receiver distances of 1200 km, beyond which the regional phases
 539 used to determine the amplitude dependence may no longer be appropri-
 540 ate. Therefore, x_{max} is limited to a maximum value of 1200 km. Second,
 541 the detection range is converted into the fractional area of Mars f_a using

542 geometry:

$$f_a(D) = \frac{1}{2} \left[1 - \cos \left(\frac{x_{max}(D)}{r_{mars}} \right) \right] \quad (19)$$

543 where r_{mars} is the radius of Mars (Figure 7b). Finally the detectable frac-
 544 tion is multiplied by the crater production functions (Figure 7c) to give the
 545 detection rate $N_{det}(D)$ for each crater diameter bin:

$$N_{det}(D) = f_a(D)N(D) \quad (20)$$

546 The resulting detection rates are plotted in Figure 7d for the two impact
 547 models. The total number of regional impacts detected is given by the sum
 548 of $N_{det}(D)$ over all crater diameters. For impact model 1 there is nominally
 549 1 detectable impact per year with a $1-\sigma$ range of 0.1-10 year⁻¹, whereas for
 550 impact model 2 there are nominally 3 detectable impacts per year with a $1-\sigma$
 551 range of 0.3-30 year⁻¹. The most commonly detected impacts are expected
 552 to have crater diameters of 5–20 m (impact model 1) and 0.5–20 m (impact
 553 model 2).

554 5. Discussion

555 Overall, I predict around 1–3 regional impacts per year will be detectable
 556 by SEIS-SP with a $1-\sigma$ uncertainty range of 0.1–30 year⁻¹.

557 The primary source of the order of magnitude uncertainty is scatter in the
 558 measured impact generated peak seismogram amplitudes, which originates
 559 from variations in crustal properties, data quality, and seismic efficiency. The
 560 scatter in both impact and explosion datasets illustrates the high variability
 561 possible in seismic coupling, even within similar terrains such as for the EA-
 562 GLE experiment. A fundamental limitation is the uncertainty and variability

563 in seismic efficiency, which depends on individual site and impact conditions.
564 It is reasonable to expect similar variations and subsequent uncertainties on
565 Mars. A secondary source of uncertainty is the size frequency distribution
566 of impact crater generation, which depends critically on the diameter where
567 atmospheric ablation, deceleration, and fragmentation become important.
568 However, this effect is not as important as might be expected; the two end
569 member impact models only change the number of detectable events by a
570 factor of ~ 3 . This is because smaller craters, which are the most heavily
571 affected by the atmosphere, are only detectable over a very limited range.

572 If we are optimistic and regard the low amplitudes from the Carancas
573 event as anomalous and the Apollo and White Sands missile impact results as
574 more representative of Mars' expected seismogram amplitudes, then around
575 10–30 regionally detectable events per year are predicted. This is tempting
576 as the Apollo / White Sands results define a consistent trend with a simi-
577 lar distance dependence to the larger and more reliable terrestrial explosion
578 dataset. However, given the sparse nature of the impact dataset, rejecting
579 any of the datapoints is not advisable.

580 The analysis presented here assumes the impact and explosion data used
581 to develop the distance-yield-amplitude scaling relation are a reasonable ana-
582 logue for determining seismic amplitudes on Mars. The validity of this ap-
583 proach depends on the attenuation and scattering properties of Mars' crust
584 and upper mantle, and how these compare to the Earth and Moon. For ex-
585 ample, the lunar regolith is highly fractured and gardened, with very high
586 scattering. It is also very dry with very low seismic attenuation (high Q).
587 While Mars' bulk attenuation has been determined from the secular acceler-

588 ation of Phobos (Smith and Born, 1976; Zharkov and Gudkova, 1997), it is
589 not possible to uniquely extract crustal and upper mantle Q values.

590 Lognonne and Mosser (1993) discuss the potential attenuation of Mars’
591 crust and mantle and conclude that martian mantle has a Q value between
592 Earth’s upper and lower mantle. Lognonne and Mosser (1993) also argue
593 that Mars’ crust should be less attenuating than Earth’s because of enhanced
594 removal of trapped fluids from crustal rocks due to Mars’ low atmospheric
595 surface pressure. However, Mars’ crust should be more attenuating than the
596 Moon, where exposure to a hard vacuum will have removed the majority of
597 trapped fluids leading to very low attenuation (high Q).

598 Crustal scattering on Mars should be present at greater levels than on
599 Earth because of the influence of the large number of impact craters, which
600 fracture and brecciate the upper crustal layers (Lognonne and Mosser, 1993).
601 However, Mars’ increased gravity and more geologically active surface should
602 mean that scattering is far less important than on the Moon. In summary,
603 we might expect attenuation and scattering for Mars to lie somewhere be-
604 tween Earth and Moon end members. Therefore, as the fitted distance-yield-
605 amplitude scaling law spans lunar and terrestrial impacts, it should provide a
606 reasonable approximation to the seismic behaviour of Mars’ crust and upper
607 mantle.

608 Despite the large uncertainties, it is interesting to compare the regional
609 predictions to global modelling results from Teanby and Wookey (2011), who
610 considered the frequency range 0.4–4 Hz. For a nominal SEIS-VBB noise level
611 of $10^{-9} \text{ ms}^{-2}\text{Hz}^{-1/2}$, Teanby and Wookey (2011) predicted that ≈ 1 event per
612 year would be detectable at 1000 km range or more, and that ≈ 0.1 events

613 per year would be detectable globally, with an order of magnitude uncertainty.
 614 For a noise level of $10^{-8} \text{ ms}^{-2}\text{Hz}^{-1/2}$, which is relevant for the SEIS-SP, these
 615 estimates became ≈ 0.1 and ≈ 0.01 events per year respectively - i.e. at the
 616 lower end of the present study's uncertainty range. The results presented
 617 here are appropriate for regional source-receiver distances of ≈ 1200 km or
 618 less, whereas the modelling results of Teanby and Wookey (2011) are valid
 619 for teleseismic source-receiver distances over 1000 km, which means the scal-
 620 ing relation results must be extrapolated somewhat to effectively compare
 621 the studies. This extrapolation is indicated by the grey lines in Figure 7.
 622 Figure 7d shows that both sets of results are only just consistent to within
 623 the errors of each study. However, much better agreement is obtained if the
 624 modelling in Teanby and Wookey (2011) is repeated using a seismic efficiency
 625 of $k_s = 5 \times 10^{-4}$ instead of $k_s = 2 \times 10^{-5}$ that was used in the original study (Fig-
 626 ure 7d). A value of $k_s = 5 \times 10^{-4}$ is roughly consistent with the upper end of
 627 laboratory studies and modelling (Güldemeister et al., 2013; Richardson and
 628 Kedar, 2013; Richardson et al., 2005; Schultz and Gault, 1975) and thus may
 629 be more appropriate for impact processes.

630 It is possible that the actual noise spectral density of SEIS-SP could be
 631 somewhat different to that specified in the mission requirements. Therefore, I
 632 have repeated the analysis in Section 4 with a range of noise levels from 10^{-9} –
 633 $10^{-7} \text{ ms}^{-2}\text{Hz}^{-1/2}$. The number of detections in each case are summarised in
 634 Table 4, including the extrapolation for impacts beyond 1200 km. Note
 635 that for instrument noise spectral densities below $\sim 10^{-9} \text{ ms}^{-2}\text{Hz}^{-1/2}$ the
 636 ambient noise is likely to be the dominant noise source. Noise on Mars is
 637 primarily determined by the wind and is expected to vary between 10^{-10} –

638 $10^{-8} \text{ ms}^{-2}\text{Hz}^{-1/2}$ (Lognonne and Mosser, 1993), depending on the time of
639 day and season.

640 Atmospheric fragmentation of meteoroids before impact could also affect
641 detection rates. Both Malin et al. (2006) and Daubar et al. (2013) observed
642 clusters of craters at the new impact sites indicating that fragmentation had
643 occurred. In the Daubar et al. (2013) study this occurred at 56% of sites.
644 A cluster of impacts will give a more complex and lower amplitude seismic
645 signal than a single large impact (Banks et al., 2015). To investigate the
646 effects of this I consider a worst case scenario where the amplitude of the
647 seismic signal generated would be determined by the largest fragment only.
648 Williams et al. (2014) modelled the fragmentation process, which included
649 a meteoroid strength parameter that was adjusted to match fragmentation
650 rates observed by Daubar et al. (2013). I consider Williams et al. (2014)’s
651 high fragmentation case and use this to determine the size-frequency dis-
652 tribution (SFD) of individual craters from Daubar et al. (2013)’s reported
653 effective diameters for the new impact sites. First, I convert the incremental
654 Daubar et al. (2013) SFD into a cumulative SFD to remove the dependence
655 on bin width. Second, I correct for underestimation bias in D_{eff} caused by
656 deceleration and ablation using the relation in Williams et al. (2014) (their
657 Figure 8b). Third, the corrected SFD is modified using the ratio of frag-
658 mentation to no-fragmentation crater production from Williams et al. (2014)
659 (their Figure 7b). Finally, the cumulative SFD is converted back into an
660 incremental SFD. Figure 8 compares the SFD of effective crater diameters
661 from Daubar et al. (2013) with the corresponding distribution of individual
662 craters assuming the fragmentation model of Williams et al. (2014). For

663 craters in the 2–40 m diameter range, fragmentation reduces the number of
664 individual craters by a factor of 0.7 compared to the SFD of effective crater
665 diameters. Therefore, while the effect of fragmentation is important, it is rel-
666 atively small compared to the order of magnitude uncertainties introduced
667 by the distance-yield-amplitude scaling relation.

668 Finally, the results presented here relate to the detectability of first arrival
669 Pn and Pg phases. These are high frequency phases suitable for detection
670 with SEIS-SP in the 1–16 Hz bandwidth. However, both SEIS-SP and SEIS-
671 VBB have sensitivity at lower frequencies (<1 Hz) where other later arriving
672 phases may be observed. In particular, the SEIS-VBB will be able to detect
673 long-period surface waves, which often have higher amplitudes than the first
674 arrivals (e.g. Benz et al., 1997; Chun and Henderson, 2009). This suggests
675 that if the Pn or Pg phase is detectable then the surface waves should also
676 be observed. Conversely, there should also be some events for which only the
677 surface waves are detectable, although these will be much more challenging
678 to interpret.

679 6. Conclusions

680 In this study I estimate the number of meteorite impacts detectable on
681 Mars with the InSight SEIS-SP instrument to be $0.1\text{--}30\text{ year}^{-1}$ with a nom-
682 inal detection rate of $1\text{--}3\text{ year}^{-1}$. These detection rates are appropriate for
683 Pn and Pg phases on regional scales and assume a nominal instrument noise
684 of $10^{-8}\text{ ms}^{-2}\text{Hz}^{-1/2}$ and a frequency bandpass of 1–16 Hz. Seismic data from
685 impacts and explosions were used to determine an empirical scaling relation
686 between peak ground velocity, source-receiver distance, and impact yield.

687 Comparison of explosion and impact datasets showed that buried explosives
 688 are ~ 10 times more efficient at generating seismic waves than surface impacts.
 689 The available impact dataset is quite limited and scatter in the measured am-
 690 plitudes caused by variations in seismic efficiency and crustal properties is
 691 the major source of uncertainty in this study. A secondary source of un-
 692 certainty is knowledge of the current crater production function on Mars.
 693 Two impact models were tested, compiled from observational sources (Ma-
 694 lin et al., 2006; Daubar et al., 2013), modelling (Williams et al., 2014), and
 695 standard isochrons (Hartmann, 2005). Choice of impact model changed the
 696 predictions by a factor of ~ 3 . An additional minor source of error was the
 697 effect of atmospheric fragmentation.

698 Comparison with the modelling study of Teanby and Wookey (2011) is
 699 possible for intermediate sized craters with 20–80 m diameters at source-
 700 receiver distances of around 1000–3000 km, which suggests that a seismic
 701 efficiency of $\sim 5 \times 10^{-4}$ may be appropriate for impact processes. This is con-
 702 sistent with laboratory and modelling results and is more optimistic than the
 703 value of 2×10^{-5} originally used by Teanby and Wookey (2011). If $k_s = 5 \times 10^{-4}$
 704 is appropriate then the estimates of Teanby and Wookey (2011) can be re-
 705 visited upwards from ≈ 0.1 to ≈ 1 globally detectable impact events per year
 706 assuming a noise of $10^{-9} \text{ ms}^{-2} \text{ Hz}^{-1/2}$ for the SEIS-VBB. However, seismic
 707 efficiency remains a major source of uncertainty in this work and further
 708 laboratory and field investigation is required.

709 For the nominal detection rate ($1\text{--}3 \text{ year}^{-1}$) or at the more optimistic
 710 end of the uncertainties ($10\text{--}30 \text{ year}^{-1}$), regional impacts should provide a
 711 viable way to study the crust and upper mantle of Mars, especially if the new

712 impact craters are locatable from orbit. Seismic recordings of impact events
713 will be complementary to any fault generated seismicity, providing different
714 frequency content, more uniform spatial distribution, and the potential for
715 accurately located events using orbital imagery. A single located impact
716 could begin to constrain crust and upper mantle velocities and the seismic
717 efficiency of the cratering process. For 5–10 impacts detected at a range of
718 distances, crude record sections could be constructed and used to identify
719 more complex seismic phases. Such a dataset could also be used to more
720 fully constrain seismic efficiency and study current cratering rates on Mars.

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D (m)	D ₁ (m)	D ₂ (m)	Impact Model 1				Impact Model 2			
			N (km ⁻² yr ⁻¹)	N _{min} (km ⁻² yr ⁻¹)	N _{max} (km ⁻² yr ⁻¹)		N (km ⁻² yr ⁻¹)	N _{min} (km ⁻² yr ⁻¹)	N _{max} (km ⁻² yr ⁻¹)	
0.0514	0.0432	0.0611	-	-	-		6.41×10^{-07}	3.21×10^{-07}	1.28×10^{-06}	‡
0.0725	0.0610	0.0863	-	-	-		3.86×10^{-06}	1.93×10^{-06}	7.72×10^{-06}	‡
0.103	0.086	0.122	-	-	-		1.46×10^{-05}	7.28×10^{-06}	2.91×10^{-05}	‡
0.145	0.122	0.173	-	-	-		4.71×10^{-05}	2.36×10^{-05}	9.43×10^{-05}	‡
0.206	0.173	0.245	-	-	-		1.02×10^{-04}	5.09×10^{-05}	2.04×10^{-04}	‡
0.290	0.244	0.345	-	-	-		1.25×10^{-04}	6.23×10^{-05}	2.49×10^{-04}	‡
0.411	0.345	0.488	-	-	-		9.53×10^{-05}	4.77×10^{-05}	1.91×10^{-04}	‡
0.580	0.488	0.690	-	-	-		5.66×10^{-05}	2.83×10^{-05}	1.13×10^{-04}	‡
0.822	0.691	0.977	-	-	-		3.18×10^{-05}	1.59×10^{-05}	6.35×10^{-05}	‡
1.162	0.977	1.382	-	-	-		1.85×10^{-05}	9.27×10^{-06}	3.71×10^{-05}	‡
1.64	1.38	1.95	5.07×10^{-08}	2.54×10^{-08}	1.01×10^{-07}	*	1.10×10^{-05}	5.52×10^{-06}	2.21×10^{-05}	‡
2.32	1.95	2.76	1.52×10^{-07}	9.65×10^{-08}	2.40×10^{-07}	*	6.26×10^{-06}	3.13×10^{-06}	1.25×10^{-05}	‡
3.28	2.76	3.90	3.55×10^{-07}	2.58×10^{-07}	4.89×10^{-07}	*	3.23×10^{-06}	1.62×10^{-06}	6.47×10^{-06}	‡
4.65	3.91	5.53	6.09×10^{-07}	4.72×10^{-07}	7.84×10^{-07}	*	1.55×10^{-06}	7.77×10^{-07}	3.11×10^{-06}	‡
6.56	5.52	7.81	5.07×10^{-07}	3.85×10^{-07}	6.68×10^{-07}	*	7.39×10^{-07}	3.69×10^{-07}	1.48×10^{-06}	‡
9.29	7.81	11.05	2.54×10^{-07}	1.75×10^{-07}	3.67×10^{-07}	*	3.48×10^{-07}	1.74×10^{-07}	6.95×10^{-07}	‡
13.08	11.00	15.56	1.53×10^{-07}	7.63×10^{-08}	3.05×10^{-07}	†	1.56×10^{-07}	7.78×10^{-08}	3.11×10^{-07}	‡
18.57	15.61	22.08	6.37×10^{-08}	3.18×10^{-08}	1.27×10^{-07}	†	6.37×10^{-08}	3.18×10^{-08}	1.27×10^{-07}	†
26.26	22.08	31.23	2.22×10^{-08}	1.11×10^{-08}	4.44×10^{-08}	†	2.22×10^{-08}	1.11×10^{-08}	4.44×10^{-08}	†
37.14	31.23	44.16	8.00×10^{-09}	4.00×10^{-09}	1.60×10^{-08}	†	8.00×10^{-09}	4.00×10^{-09}	1.60×10^{-08}	†
52.56	44.20	62.50	3.15×10^{-09}	1.57×10^{-09}	6.29×10^{-09}	†	3.15×10^{-09}	1.57×10^{-09}	6.29×10^{-09}	†
74.29	62.47	88.34	1.10×10^{-09}	5.50×10^{-10}	2.20×10^{-09}	†	1.10×10^{-09}	5.50×10^{-10}	2.20×10^{-09}	†
105.06	88.34	124.94	4.07×10^{-10}	2.03×10^{-10}	8.13×10^{-10}	†	4.07×10^{-10}	2.03×10^{-10}	8.13×10^{-10}	†
148.32	124.73	176.39	1.46×10^{-10}	7.28×10^{-11}	2.91×10^{-10}	†	1.46×10^{-10}	7.28×10^{-11}	2.91×10^{-10}	†

Table 1: The two models of present day cratering rates used in this study. Impact Model 1 is based on observed new small craters from Daubar et al. (2013) and Malin et al. (2006) combined with a rescaling of the 1 Gyr isochron of Hartmann (2005) for larger crater diameters and should be considered a lower bound on current impact rate. Impact Model 2 additionally incorporates modelling of smaller sub-observation scale impactors from Williams et al. (2014). Columns are: D crater diameter bin centre; D_1/D_2 minimum/maximum limits of crater diameter bin; N incremental cratering rate for each bin; $N_{\min}/N_{\max}$ minimum/maximum cratering rate including all error contributions. Sources: *Daubar et al. (2013), error bar from Poisson statistics; † Hartmann (2005) 1 Gyr isochron scaled by $1/3 \times 10^{-9}$ to match Malin et al. (2006) and Daubar et al. (2013) observations, error bar standard factor of 2 error discussed in Hartmann (1999, 2005); ‡ modelling results from Williams et al. (2014) scaled to match the Daubar et al. (2013) crater counts in the 9.29 and 13.08 m bins. Bins are spaced by a factor of $\sqrt{2}$ and have the same bin centres as Hartmann (2005). Values are plotted in Figure 1.

Type	Source	Date (UT)	Time (UT)	Longitude (°E)	Latitude (°N)	Depth (m)	Yield (kg TNT)
Impact	Carancas Meteor Impact	15/09/2007	16:40:14	-69.044	-16.664	0.0	2000
	Apollo 14 LM	07/02/1971	00:45:25	-19.670	-3.420	0.0	778
	Apollo 16 SIVB	19/04/1972	21:02:04	-23.800	1.300	0.0	11000
	Apollo 17 SIVB	10/12/1972	20:32:42	-12.310	-4.210	0.0	11300
Explosion	EAGLE, SP11, Goha Tsion	11/01/2003	21:20:01	38.285	9.982	50.0	900
	EAGLE, SP12, Gerba Guracha	12/01/2003	21:20:00	38.476	9.759	50.0	1900
	EAGLE, SP13, Derba	13/01/2003	21:20:05	38.674	9.361	50.0	600
	EAGLE, SP14, Cheffe Donsa	12/01/2003	21:03:00	39.121	8.980	51.6	375
	EAGLE, SP16, Kula	13/01/2003	21:10:00	39.688	8.019	50.0	1100
	EAGLE, SP17, Bele	12/01/2003	21:10:00	40.032	7.727	50.0	2500
	EAGLE, SP18, Delo Sebro	11/01/2003	21:29:55	40.515	7.233	50.0	2200
	EAGLE, SP24, Koka	13/01/2003	21:50:00	39.005	8.328	51.5	525
	EAGLE, SP25, Doni B	12/01/2003	21:50:00	39.618	8.532	70.0	1025
	EAGLE, SP26, Beseka	14/01/2003	21:40:00	39.860	8.829	50.0	1150
	EAGLE, SP28, Gewane	11/01/2003	21:40:00	40.641	10.214	50.0	1000
	UK Nuclear Test, “Bristol”	26/11/1991	18:35:00	-116.070	37.096	457.0	11 000 000
	US Nuclear Test, “Divider”	23/09/1992	15:04:00	-115.989	37.021	426.0	5 000 000
	China Nuclear Test	29/07/1996	01:48:57	88.420	41.820	-	3 000 000
	DPRK Nuclear Test 1	09/10/2006	01:35:28	129.108	41.287	-	480 000
	DPRK Nuclear Test 3	12/02/2013	02:57:51	129.076	41.291	-	12 200 000

Table 2: Impact and explosive source parameters. Data compiled from: Carancas Impact (Brown et al., 2008; Le Pichon et al., 2008; Tancredi et al., 2009); Apollo Impacts (Toksoz et al., 1974; Williams, 2003); EAGLE chemical explosives (Maguire, 2003); UK/US nuclear tests (U. S. Department of Energy, 2000); Chinese nuclear test (Yang et al., 2003; CTBTO, 2012); and DPRK nuclear tests (Zhang and Wen, 2013).

Parameter	Value	Fractional Error	Notes
a_0	1.825×10^{-5}	2.45	Value for Earth fitted to explosion data in Figure 6
b	-1.60	0.023	Fitted to explosion data in Figure 6
c	0.5	-	Fixed (0.49 ± 0.03 if fitted to data in Figure 6)
q	$4.18 \times 10^6 \text{ J kg}^{-1}$	-	TNT yield (Shoemaker, 1983)
s	0.099	3.82	Fitted to impact data in Figure 6
a	5.568×10^{-5}	2.45	Value for Mars (eqn. 7)
$\alpha_{\oplus}$	8.8×10^{-3}	0.35	Teanby and Wookey (2011)
β	0.32	0.03	Teanby and Wookey (2011)
α	1.06×10^{-2}	0.35	Scaled to Mars gravity using eqn. 10
p_a	$10^{-8} \text{ ms}^{-2} \text{ Hz}^{-1/2}$	-	SEIS-SP power spectral density noise requirement
f_1, f_2	1, 16 Hz	-	Frequency range of regional events
f	4 Hz	-	Nominal frequency
p_v	$4.0 \times 10^{-10} \text{ ms}^{-1} \text{ Hz}^{-1/2}$	-	From eqn. 13 and p_a
n_v	$1.9 \times 10^{-9} \text{ ms}^{-1}$	-	Peak velocity noise from eqn. 14
$g_{\oplus}$	9.81 ms^{-2}	-	Earth gravity
g	3.71 ms^{-2}	-	Mars gravity
r_{mars}	3 392 000 m	-	Mars radius

Table 3: Numerical values of parameters used to determine detectability of regional impacts. For impacts the overall uncertainty is dominated by the large errors in s and α .

Impact Model 1

Seismometer	Regional Impacts (<1200 km)				All Impacts (extrapolated)			
Noise	N_{det}	1σ range			N_{det}	1σ range		
($\text{ms}^{-2}\text{Hz}^{-1/2}$)	(yr^{-1})	(yr^{-1})			(yr^{-1})	(yr^{-1})		
1×10^{-7}	0.065	0.0055	–	0.77	0.065	0.0055	–	0.77
3×10^{-8}	0.28	0.025	–	3.3	0.29	0.025	–	3.4
1×10^{-8}	1.0	0.095	–	11	1.1	0.097	–	13
3×10^{-9}	3.3	0.39	–	34	5.0	0.43	–	57
1×10^{-9}	7.1	1.2	–	58	18	1.7	–	190

Impact Model 2

Seismometer	Regional Impacts (<1200 km)				All Impacts (extrapolated)			
Noise	N_{det}	1σ range			N_{det}	1σ range		
($\text{ms}^{-2}\text{Hz}^{-1/2}$)	(yr^{-1})	(yr^{-1})			(yr^{-1})	(yr^{-1})		
1×10^{-7}	0.17	0.015	–	2.1	0.17	0.015	–	2.1
3×10^{-8}	0.78	0.066	–	9.2	0.78	0.066	–	9.3
1×10^{-8}	2.9	0.26	–	34	3.1	0.26	–	37
3×10^{-9}	12	1.1	–	140	14	1.2	–	160
1×10^{-9}	40	4.1	–	450	52	4.6	–	600

Table 4: Number of detectable impacts N_{det} as a function of acceleration noise spectral density in the 1–16 Hz bandpass for the two impact models. A noise of $1 \times 10^{-8} \text{ ms}^{-2}\text{Hz}^{-1/2}$ is the nominal SEIS-SP specification. For this and higher noise levels the majority of detectable impacts are regional, so there is minimal difference between the number of regional impacts detected and the extrapolated total number of impacts detected. For lower noise levels, impacts further away than 1200 km begin to make up a significant proportion of the detectable events.

Figure 1: Current crater production function models and observations. $N(D)$ is the incremental number of craters in a bin centred on crater diameter D , with range $2^{-1/4}D$ to $2^{1/4}D$ (per km^2 on the left axis and for the whole of Mars for the right axis). Observations are from Malin et al. (2006) and Daubar et al. (2013). The downturn in $N(D)$ at small crater diameters in the new crater observations is attributed to finite image resolution or atmospheric ablation and deceleration. Values for the impact model curves are given in Table 1.

Figure 2: Location maps of impact and explosion datasets used to determine the distance-yield-amplitude scaling relation. Red circles are events (impacts or explosions) and blue triangles are seismometers. (a) Apollo artificial lunar impacts overlain on the Clementine lunar basemap. (b) Carancas impact event. (c) EAGLE controlled source chemical explosions. (d) US nuclear tests at the Nevada test site. (e) Chinese nuclear test. (f) North Korean nuclear tests.

Figure 3: Seismic data from Apollo artificial impacts. (a) Apollo 14 lunar module (LM) impact recorded on the Apollo 14 seismometer. (b) Apollo 16 Saturn V booster stage (SIVB) recorded on Apollo 14 and 15 seismometers. (c) Apollo 17 SIVB recorded on Apollo 14, 15, and 16 seismometers. Vertical dashed line at 0 seconds indicates the impact time. Light grey region indicates noise window and dark grey window indicates first arrival window. Seismograms are from the short-period vertical sensor after deconvolution of the instrument response. Source-receiver distances shown on right of plot. Note the emergent nature of seismic events makes identification of a distinct first arrival phase difficult. Therefore, I used 20 seconds after the first arrival onset as a nominal first arrival window. Seismograms have been bandpass filtered between 1–16 Hz using a 4 pole Butterworth filter.

Figure 4: Seismic data from the Carancas impact event on 15th September 2007. Vertical dashed lines indicate impact origin time (16:40:14 UT). Light grey region indicates noise window and dark grey window indicates first arrival window. (a) Short-period record from BSN station BOD. High amplitude arrival at >75 s is the airwave. (b,c) Short (SHZ) and long (BHZ) period records at GSN station LPAZ. Both instruments give comparable first arrival amplitudes. (d) short-period recording at BSN station BOE. All seismograms have been bandpass filtered using a 1–16 Hz 4 pole Butterworth filter.

Figure 5: Example record sections from explosive data. (a) EAGLE shot point 14 at Cheffe Donsa, (b) US nuclear test “Divider”, (c) Chinese 1996 nuclear test, and (d) DPRK 2006 nuclear test. Note that the EAGLE data in (a) is shown in reduced time relative to explosion origin time for clarity, whereas nuclear test data in (b,c,d) are in time relative to the reported explosion origin time. Light grey region indicates noise window and dark grey window indicates first arrival window. All seismograms have been bandpass filtered using a 1–16 Hz 4 pole Butterworth filter.

Figure 6: Peak seismogram amplitudes as a function of source-receiver distance from explosive and impact datasets. Solid lines show lines of best fit and dashed lines indicate 1σ uncertainties due to data scatter. Scaled velocity is the peak seismogram velocity in the first arrival window scaled by the square root of the yield as in equation 3 such that it is the equivalent peak seismogram velocity for a 1000 kg TNT event.

Figure 7: Detectability of regional impacts on Mars. (a) Maximum detection range of regional impacts as a function of crater diameter (black solid line) and associated 1σ uncertainties (black dashed lines). Impacts are only considered regional if they are within 1200 km of the seismometer, which is the range over which the peak amplitude scaling was determined. Hence, the detection range has a its maximum value set to 1200 km. Grey lines indicate an extrapolation of the scaling relation to greater source-receiver distances. (b) Detection range expressed as a proportion of Mars' surface. (c) Current crater production functions for impact models 1 and 2 along with 1σ uncertainties (dashed lines). (d) Product of (b) and (c), which gives the number of detectable regional events (range <1200 km) per $\sqrt{2}$ diameter bin for impact model 1 (black circles) and impact model 2 (open diamonds). Dashed lines show 1σ uncertainties for each model, which are dominated by the error in s (Table 3). Grey points indicate number of detections at all distances using the extrapolated scaling relation. Open square at 74.29 m shows the prediction from Teanby and Wookey (2011) who assume a seismic efficiency $k_s = 2 \times 10^{-5}$. Open inverted triangles show re-calculated predictions from Teanby and Wookey (2011) assuming $k_s = 5 \times 10^{-4}$. This higher value of k_s gives much better agreement with predictions from the extrapolated scaling law. Note the model predictions from Teanby and Wookey (2011) also have an order of magnitude uncertainty, which is not shown for clarity.

Figure 8: Effect of meteoroid fragmentation on number of observed craters. (a) Black circles with error-bars show the number of craters in each $\sqrt{2}$ crater diameter bin observed by Daubar et al. (2013), where clusters of craters assumed to be from the same meteoroid have been combined into a single effective crater diameter. Open circles show the number of individual craters predicted by mapping the observed crater numbers through the high fragmentation model of Williams et al. (2014). (b) Fragmentation causes a reduction in the formation rates of individual craters in the 2–40 m diameter range by a factor of ≈ 0.7 . The effect of fragmentation is thus small compared to other error sources in this study.