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1 **Simulation of topography effects on rockfall-generated**
 2 **seismic signals: application to Piton de la Fournaise**
 3 **volcano**

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16 **Key Points:**

- 17 • First-ever simulation of high-frequency rockfall seismic waves using the 3D Spec-
 18 tral Element Method
- 19 • Ground-motion amplification induced by volcano topography found to be depen-
 20 dent on soil properties and rockfall position
- 21 • Simulations and observations successfully compared by means of inter-station spec-
 22 tral ratios and Hertz theory

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23 **Abstract**

24 Seismic waves generated by rockfalls contain valuable information on the properties of
 25 these events. However, as rockfalls mainly occur in mountainous regions, the generated
 26 seismic waves can be affected by strong surface topography variations. We present a method-
 27 ology for investigating the influence of topography using a Spectral-Element-based sim-
 28 ulation of 3D wave propagation in various geological media. This methodology is applied
 29 here to Dolomieu crater on the Piton de la Fournaise volcano, Reunion Island, but it can
 30 be used for other sites, taking into account local topography and medium properties.

31 The complexity of wave fields generated by single-point forces is analyzed for dif-
 32 ferent velocity models and topographies. Ground-motion amplification is studied rela-
 33 tive to flat reference models, showing that Peak Ground Velocity (PGV) and total ki-
 34 netic energy can be amplified by factors of up to 10 and 20, respectively. Simulations
 35 with Dolomieu-like crater shapes suggest that curvature variations are more influential
 36 than depth variations.

37 Topographic effects on seismic signals from rockfalls at Dolomieu crater are revealed
 38 by inter-station spectral ratios. Results suggest that propagation along the topography
 39 rather than source direction dominates the spectral ratios and that resulting radiation
 40 patterns can be neglected.

41 The seismic signature of single rockfall impacts is studied. Using Hertz contact the-
 42 ory, impact force and duration are estimated and then used to scale simulations, achiev-
 43 ing order-of-magnitude agreement with observed signal amplitudes and frequency thresh-
 44 olds. Our study shows that combining Hertz theory with high-frequency seismic wave
 45 simulations on real topography improves the quantitative analysis of rockfall seismic sig-
 46 nals.

47 **1 Introduction**

48 Interactions between seismic wave fields and complex surface geometries can locally
 49 modify seismic ground motion. Anomalously strong shaking on hilltops and mountain
 50 ridges or flanks, often causing severe structural damage to buildings (W. H. K. Lee et
 51 al., 1994; Hartzell et al., 1994; Hough et al., 2010) or triggering earthquake-induced land-
 52 slides (Meunier et al., 2008; Harp et al., 2014), have been related to seismic amplifica-
 53 tion due to such topographic effects. Data from field experiments support the assump-
 54 tion of amplified ground motion at the top compared to the bottom of a mountain (Davis
 55 & West, 1973; Pedersen et al., 1994; Spudich et al., 1996).

56 Numerous studies have tried to quantify numerically the topographic effect on seis-
 57 mic waves generated by deep sources. Geli et al. (1988) provided an extensive review of
 58 previous studies together with new results from more complex models (i.e. including sub-
 59 surface layering and neighboring ridges). Using an earthquake simulation with three-dimensional
 60 topography, Bouchon and Barker (1996) found that a small hill of less than 20-m high
 61 can amplify ground acceleration by 30% to 40% for frequencies between 2 Hz and 15 Hz.
 62 Using the 3D spectral element method, S. J. Lee, Chan, et al. (2009) studied the effects
 63 of high-resolution surface topography. They found that values of Peak Ground Accel-
 64 eration (PGA) can be increased up to 100% relative to simulations on a flat surface and
 65 reported an increase in cumulative kinetic energy of up to 200% as a result of increased
 66 duration of shaking linked to complex reflection and scattering processes during the in-
 67 teraction of the seismic waves with the topography.

68 Yet, because of complex patterns of amplification and deamplification, it is diffi-
 69 cult to quantify the effect of topography in a generic way. Maufroy et al. (2015) proposed
 70 to use the topography curvature, smoothed over a characteristic length depending on the
 71 studied wavelength, as a proxy for amplification factors. Based on the NGA-West2 earth-
 72 quake catalog (Ancheta et al., 2014), Rai et al. (2017) showed statistical biases of site
 73 residuals in the ground-motion prediction equation (GMPE, Chiou & Youngs, 2014) to-
 74 wards relative elevation and smoothed curvature and suggested topographic modifica-
 75 tion factors dependent on signal frequency and relative elevation. In addition to these

76 successful findings, other authors have pointed out the complex coupling between topog-
 77 raphy and the underlying soil structure that must not be neglected when estimating to-
 78 pographic amplification (Assimaki & Jeong, 2013; Hailemichael et al., 2016; B. Wang et
 79 al., 2018; Jeong et al., 2019).

80 All the studies mentioned above investigate topographic effects on a seismic wave
 81 field of vertical incidence. S. J. Lee, Komatitsch, et al. (2009) investigates the influence
 82 of the source depth on ground motion amplification and demonstrates that amplifica-
 83 tion in a basin can be reduced when a mountain range is located between the basin and
 84 a shallow source. This suggests that surface topography can have a pronounced influ-
 85 ence on the propagation of surface waves subjected to an accumulated effect of scatter-
 86 ing, diffraction, reflection, and conversion. It is crucial to enhance our understanding of
 87 these mechanisms for the study of shallow seismic sources that have gained increasing
 88 attention in the emerging field of environmental seismology (Larose et al., 2015). Sev-
 89 eral authors have investigated numerically the interaction of surface waves with 2D sur-
 90 face geometries such as corners, hills or canyons (Munasinghe & Farnell, 1973; Weaver,
 91 1982; Snieder, 1986; Sánchez-Sesma & Campillo, 1993; Zhang et al., 2018; B. Wang et
 92 al., 2018). Ma et al. (2007) demonstrated that a topographic feature 10 times smaller
 93 than the wavelength can still considerably reduce the amplitude of by-passing surface
 94 waves. Similar to S. J. Lee, Komatitsch, et al. (2009), they simulated the shielding ef-
 95 fects of large-scale topography on fault-generated surface waves using a 3D model of the
 96 San Gabriel Mountains, Los Angeles, California, finding amplification factors in peak ground
 97 velocity (PGV) of up to +50% on the source-side of the mountain range and up to -50%
 98 on the opposite site. L. Wang et al. (2015) modeled the influence of an uplifted and a
 99 depressed topography on the wave field. Comparing amplitudes and frequency content
 100 between source side and far-source side, they found that the depressed topography caused
 101 stronger contrasts than the uplifted topography, especially for steeper slopes and at higher
 102 frequencies.

103 The present study is focused on seismic waves generated by rockfalls. Different than
 104 the source mechanism of earthquakes, rockfall seismic sources can generally be described
 105 by impulse forces on the Earth’s surface. Seismic signals from rockfalls, or more gener-
 106 ally from landslides, have been demonstrated to be very useful to classify and locate events
 107 as well as constrain flow dynamics and rheology (e.g. Vilajosana et al., 2008; Deparis
 108 et al., 2008; Favreau et al., 2010; Hibert et al., 2011; Dammeier et al., 2011; Moretti et
 109 al., 2012; Bottelin et al., 2014). However, as landslides predominantly occur in areas of
 110 strong topographic relief, the measurements can be strongly influenced by topography
 111 variations leading to erroneous landslide estimates. For example, to calculate landslide
 112 volumes, the generated seismic energy is estimated from seismic recordings (Hibert et
 113 al., 2011). At the same time, energy estimations can vary from station to station. The
 114 present work shows that the topography studied here can partly explain amplitude vari-
 115 ations between seismic stations.

116 In the following, after introducing the study site located at Dolomieu crater on Piton
 117 de la Fournaise volcano, Reunion Island, the numerical model for the SEM simulations
 118 is defined, entailing a discussion on the seismic velocity profile at Piton de la Fournaise.
 119 As the mesh size affects the computational cost, different topography resolutions are com-
 120 pared. Then, topography induced amplification is computed, depending on the under-
 121 lying velocity model, by means of peak ground velocity (PGV) and total kinetic energy
 122 for both vertical and horizontal sources. In an attempt to quantify the dependencies on
 123 geometric parameters, the influence of variations in crater depth and curvature are in-
 124 vestigated.

125 Finally, real seismic signals generated by rockfalls at Dolomieu crater are analyzed.
 126 Simulated and observed inter-station spectral ratios are compared, making it possible
 127 to examine the spectral content of the signals independently of the rockfall source. Ad-
 128 ditionally, the seismic signature of single rockfall impacts is investigated. To compare
 129 signal amplitude and frequency content between observations and simulations, impact
 130 force and duration are estimated based on Hertz contact theory (Hertz, 1878).

2 Study site

The study site is located on Piton de la Fournaise volcano, Reunion Island, presented in Figures 1a and 1b. Its summit is characterized by the 340 m deep Dolomieu crater that collapsed in 2007 (e.g. Staudacher et al., 2009). Because of the instability of the crater walls, rockfall events are frequently observed within the crater (Hibert et al., 2011; Hibert, Mangeney, et al., 2014, 2017; Durand et al., 2018; Derrien et al., 2019).

The high quantity of events together with a dense seismic network monitored by the *Observatoire Volcanologique du Piton de La Fournaise* (OVPF) provide excellent conditions for the study of rockfalls. Using recorded seismic signals, past studies have investigated the links between rockfall activity and external forcings such as rain or seismicity, the spatio-temporal evolution of rockfall occurrences as well as their volumes (Hibert, Ekström, & Stark, 2014; Hibert, Mangeney, et al., 2017; Durand et al., 2018). In addition to the seismic stations, three cameras positioned on the crater rim continuously monitor rockfall activity. This makes it possible to correlate video images and rockfall seismic signals.

For example, Figures 1c-1e show images and seismic signals of a rockfall on the southern crater wall on February 28, 2016. The rockfall consisted of mainly three boulders that can clearly be traced on the video provided in the Supporting Information. Each boulder took around 30 s to move from the top to the bottom of the crater.

The first movement can be seen in snapshot i). At that time, a large signal amplitude was recorded on station DSO, located very close to the source position. Subsequently, the rockfall traveled through a small valley in ii) and accelerated towards the position in iii). The acceleration of the boulder resulted in strong impacts which can be seen on both the signal and the spectrogram after time iii) at all stations. At the time corresponding to snapshot iv), the first boulder arrived at the crater bottom, whereas a second boulder was half-way down. Again, strong amplitudes were measured around time iv), probably corresponding to the second boulder. Around time v), the last movements of a third block are visible. After this, residual granular activity distributed on the flank can be observed on the video. Signal amplitudes decay accordingly.

Note that station DSO recorded very strong signals in the beginning, while signal amplitudes increased slowly at the other stations. This is certainly related to the changing source-receiver distance. Additionally, topography may have influenced the signal amplitudes depending on the source position relative to the receiver position. From the spectrograms, we can see that the main frequency content was between 3 Hz and 20 Hz.

3 SEM simulations

To study the effect of topography on rockfall seismic signals recorded at different stations, the seismic wave propagation was simulated using the 3D Spectral Element Method (SEM, e.g. Festa & Vilotte, 2005; Chaljub et al., 2007). The seismic impulse response was modeled by implementing a point force at the surface of the domain in the form of a Ricker wavelet with a dominant frequency of 7 Hz, covering a bandwidth from 2 Hz to 20 Hz, which is predominantly observed for rockfalls at Dolomieu crater. The source magnitude was set to unity except for analysis of a single impact rockfall in section 5.3.

Moving rockfall source positions and poorly known subsurface properties require many simulations with different configurations. For easy reference, a table is provided in Appendix A, listing the simulation configurations used in the different sections of the article.

3.1 Mesh of the Earth model

Figure 2a shows a cross-section through the spectral-element mesh with Dolomieu surface topography. The dimensions of the domain measure $x = 2100$ m (easting), $y =$

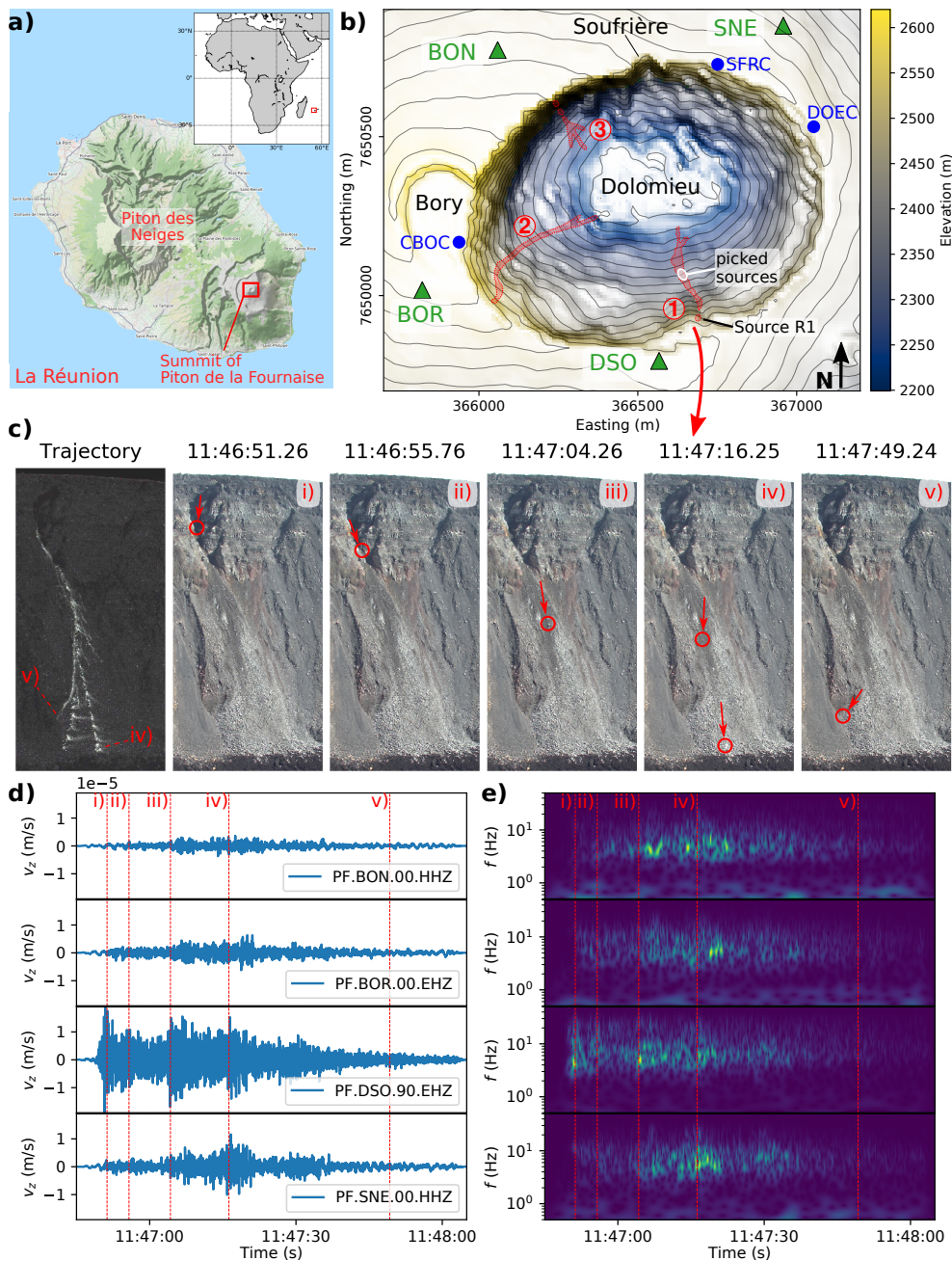


Figure 1. a) Map of Reunion Island with dormant volcano Piton des Neiges and active volcano Piton de la Fournaise. b) The summit of Piton de la Fournaise with 340 m deep Dolomieu crater and smaller craters Bory and Soufrière. Trajectories of three rockfalls are indicated by red zones. Seismic stations BON, BOR, DSO, and SNE (green triangles); cameras CBOC, DOEC, and SFRC (blue dots). Contour lines show elevation differences of 20 m. c) Trajectory and snapshots from camera SFRC of rockfall 1 at the southern crater wall on February 28, 2016. Circles and arrows mark a selection of boulder positions and their direction of arrival. A video of the rockfall is provided in the Supporting Information. d) Vertical ground velocity recorded at all four stations. Vertical lines from i) to v) mark the times of camera snapshots in c). e) Corresponding spectrograms (Stockwell transform).

181 1800 m (northing), and $z = 600$ m (depth). To simulate an open domain, 160 m thick
 182 absorbing PML boundaries (Perfectly Matched Layers, e.g. Festa & Vilotte, 2005) are
 183 added on the sides and bottom of the domain. The elements are successively deformed
 184 in the vertical direction to accommodate the topography provided by a Digital Eleva-
 185 tion Model (DEM) with a 10 m resolution. To decrease computational costs, the element
 186 size is increased from 10 m to 30 m at 150 m below the surface (*Zone of refinement*), re-
 187 sulting in a total of 915,704 elements. To filter out short wavelength variations of the
 188 fine mesh that cannot be represented in the coarse mesh, a smooth *Buffer layer*, provided
 189 by a low-pass filtered topography, is used as an additional boundary 100 m below the sur-
 190 face. Simulations are implemented using a polynomial degree of 5, i.e. 6 GLL points (Gauss-
 191 Lobatto-Legendre, e.g. Chaljub et al., 2007) per element in each direction.

192 Given the high computational costs (i.e. CPU times of up to 460 days, with 10 cores
 193 per CPU, for the heterogeneous velocity model), a mesh with a reduced topography res-
 194 olution of 20 m is used in the first part of this study in which different velocity models
 195 are explored with a fixed point source located at the southern crater wall. This mesh is
 196 built using elements with a constant side length of 20 m, which reduces the total num-
 197 ber of elements to 550,000, accordingly decreasing the CPU time to 145 days for the het-
 198 erogeneous velocity model. The reduced number of elements also increases memory ef-
 199 ficiency when displaying snapshots.

200 In the second part of the study, when comparing simulations to observations of rock-
 201 falls at Dolomieu crater, the model with high-resolution topography is used. Here we ap-
 202 ply the reciprocity principle (Bettuzzi, 2009), i.e. the synthetic source is located at the
 203 position of the real seismometers (BON, BOR, DSO and SNE) and the wave field is recorded
 204 on a 10×10 m grid of stations across Dolomieu crater. In this way, the impulse responses
 205 of all potential rockfall sources are modeled with just one simulation per seismometer
 206 and per channel.

207 The mesh of the flat-surface reference model is built with elements with 20 m side
 208 lengths. Cross-sections through all meshes (i.e. those with a flat surface; with 20 m-resolution
 209 topography; with 10 m-resolution topography) are provided in the Supporting Informa-
 210 tion.

211 3.2 Velocity model

212 Three different velocity models are used: (1) a homogeneous model, (2) a model
 213 with shallow low S-wave velocity layer, and (3) a model with smoothly increasing veloc-
 214 ity as proposed by Lesage et al. (2018) for shallow volcano structures. The velocity-depth
 215 profiles are illustrated in Figure 2b and summarized in Table 1. The generic model by
 216 Lesage et al. (2018) is based upon measurements on multiple andesitic and basaltic vol-
 217 canoes. P- and S-wave speeds c_i are expressed as follows:

$$c_i(z) = c_{i0}[(z + a_i)^{\alpha_i} - a_i^{\alpha_i} + 1], \quad i = P, S, \quad (1)$$

218 where z is the depth below the surface, c_{i0} are the velocities at zero depth, and α_i and
 219 a_i are fitting parameters as defined in Table 1.

220 The velocity profiles are compared to the S-wave velocity model inverted from am-
 221 bient noise recordings at Piton de la Fournaise by Mordret et al. (2015). The orange-
 222 shaded zone shown in Figure 2b corresponds to depth-profiles extracted from the inverted
 223 3D model in the vicinity of Dolomieu crater. Good agreement is observed with the Lesage
 224 generic velocity profile. The discrepancy in the first 100 m can be caused by missing high-
 225 frequency content in the model of Mordret et al. (2015), who inverted frequencies be-
 226 low 2.5 Hz.

227 In order to further validate the Lesage generic model for our study site, Rayleigh
 228 velocity dispersion curves from noise measurements at a mini-array located around sta-
 229 tion BON are compared in Figure 2c with theoretical dispersion curves of the Lesage generic
 230 model. Picks from the mini-array measurements are determined using the Modified Spa-
 231 tial Autocorrelation (MSPAC) Toolbox (Köhler et al., 2007; Wathelet et al., 2008) as im-

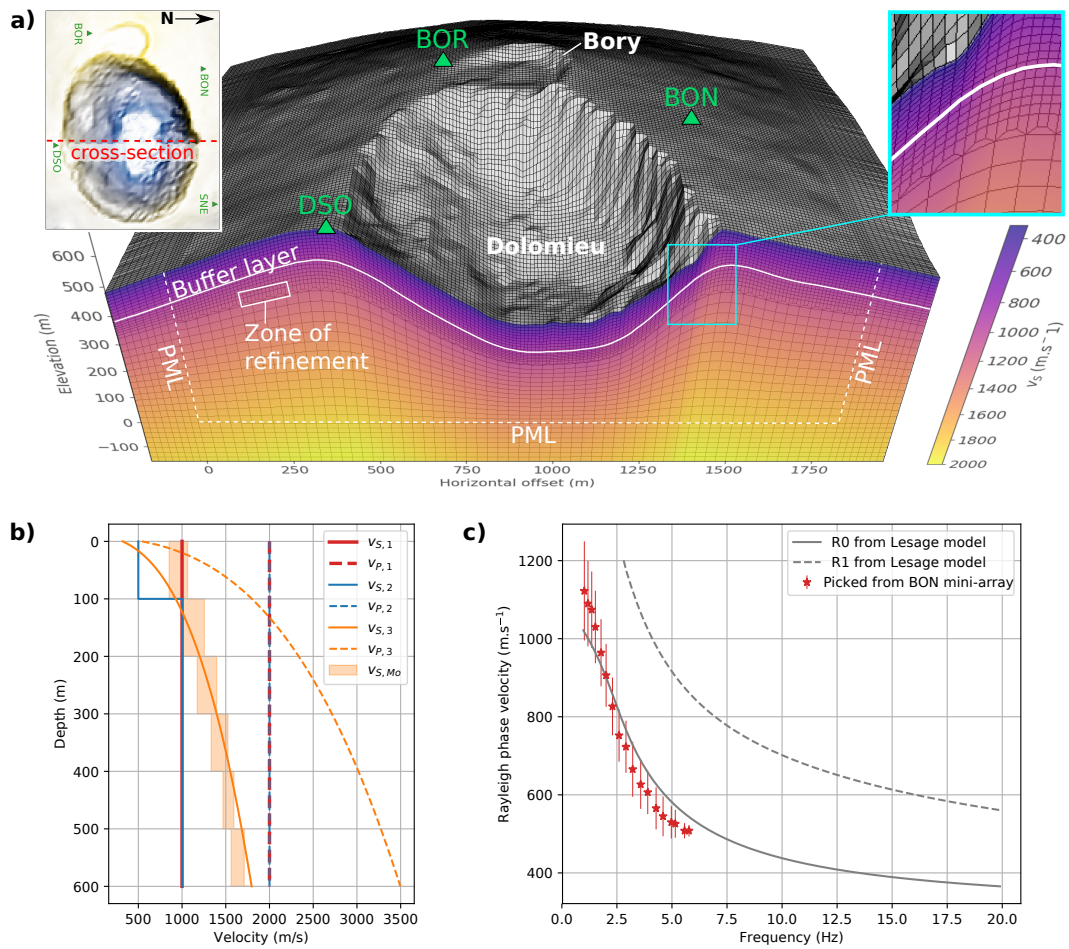


Figure 2. a) Cross-section of the SEM mesh through Dolomieu crater with a topography resolution of 10 m. Perspective as seen from the East with Bory crater located in the background. The color map corresponds to the Lesage generic velocity model (see section 3.2). The buffer layer 100 m below the surface dampens small-scale topography variations. The zone of refinement at 150 m below the surface connects elements with 10 m and 30 m side lengths. 160 m wide PML boundaries are attached to the sides and bottom of the domain. b) S- and P-wave velocity depth profiles for the (1) homogeneous model ($v_{S,1}$ and $v_{P,1}$), (2) model with shallow S-wave velocity layer ($v_{S,2}$ and $v_{P,2}$), and (3) Lesage generic velocity model ($v_{S,3}$ and $v_{P,3}$). The shaded zone ($v_{S,Mo}$) is extracted from the inverted 3D S-wave model of Mordret et al. (2015). c) Theoretical dispersion curves of the Lesage generic model for the fundamental (R0) and first-mode (R1) Rayleigh wave velocity together with picked dispersion curves from a mini-array around station BON. The errors are estimated from the uncertainty during dispersion curve picking.

232plemented in Geopsy software (www.geopsy.org). Theoretical dispersion curves are cal-
 233culated from the Lesage generic model using modal summation from *Computer Programs*
 234*in Seismology* (Herrmann, 2013). The measured values agree well with the fundamen-
 235tal Rayleigh velocity dispersion curve. No coherent dispersion curves could be picked above
 2366 Hz because of the minimum mini-array aperture of 30 m.

237Despite missing measurements above 6 Hz, the Lesage generic model is assumed
 238to be the most reasonable model for the shallow high-frequency velocity structure of Piton
 239de la Fournaise volcano because it is based upon measurements at comparable volcanoes.

Table 1. Parameters of velocity models for the SEM simulations^a

Velocity model	v_P	v_S	ρ (kg m ⁻³)	Q_P	Q_S
1) homogeneous	2000 m s ⁻¹	1000 m s ⁻¹	2000	80	50
2) low v_S layer	2000 m s ⁻¹	500 m s ⁻¹ (top 100 m) 1000 m s ⁻¹ (below 100 m)	2000	80	50
3) generic	$c_{P0} = 540$ m s ⁻¹ $\alpha_P = 0.315$ $a_P = 10$	$c_{S0} = 320$ m s ⁻¹ $\alpha_S = 0.300$ $a_S = 15$	2000	80	50

^aP- and S-wave velocity v_P and v_S , density ρ , and P- and S-wave quality factor Q_P and Q_S for the (1) homogeneous model, (2) model with shallow S-wave velocity layer, and (3) Lesage generic velocity model.

240 The Lesage generic velocity model implemented on the SEM mesh is represented
 241 in Figure 2a. There are two options when implementing a velocity-depth profile on a 3D
 242 numerical domain with topography. The first possibility is to keep the velocity laterally
 243 homogeneous and excavate a surface corresponding to the topography. The second possi-
 244 bility is to adjust the velocity profile vertically so that it follows the topography ele-
 245 vation. Either way, the subsurface velocity structure is influenced, unless it is homoge-
 246 neous. We chose the second option that we believe is geologically more reasonable be-
 247 cause a main cause of velocity variation is the compaction of material with depth due
 248 to increasing overburden pressure.

249 Rock density ρ as well as quality factors Q_P and Q_S for intrinsic attenuation of
 250 P- and S-wave velocity, respectively, are chosen based on previous studies on Piton de
 251 la Fournaise and similar volcanoes (Battaglia, 2003; O’Brien & Bean, 2009; Hibert et al.,
 252 2011). All parameters are summarized in Table 1.

253 This work focuses on the topography effect, but it is important to have an idea of
 254 the effect of 3D-medium heterogeneities. Difficulties arise in this respect as there is lack
 255 of knowledge on the distribution of heterogeneities that is hard to invert from seismo-
 256 grams alone (Imperator & Mai, 2013). Nevertheless, a first attempt to simulate scat-
 257 tering effects is made by adding a spatially random velocity perturbation to the Lesage
 258 generic velocity model. The magnitude of the velocity deviation reaches 43% and is de-
 259 fined by a normal distribution with a standard deviation of 10%, see Supporting Infor-
 260 mation.

261 3.3 Topography resolution

262 The influence of topography resolution on the simulated wave field is investigated
 263 to study influences from sub-wavelength topography variations and assess the trade-off
 264 between increased resolution and computation costs. Synthetic seismograms of the ver-
 265 tical component are compared (Figure 3a), obtained from models with a flat surface and
 266 topography resolutions of 20 m and 10 m. A vertical point force in the form of 7 Hz Ricker
 267 source-time function is placed on the southern crater wall, corresponding to rockfall start-
 268 ing position R1 in Figure 1b.

269 The single-impact source produces a long wave-train of body waves and multiple-
 270 mode Rayleigh waves. The maximum amplitudes decrease for the models with topog-
 271 raphy compared to the simulations with the flat model. The large-scale crater topog-
 272 raphy can redirect the wave-field and cause shadow-zones, as shown in section 4 where the
 273 spatial distribution of amplification or deamplification along the surface will be analyzed.
 274 Furthermore, topography causes prolonged and more complex waveforms. For the flat

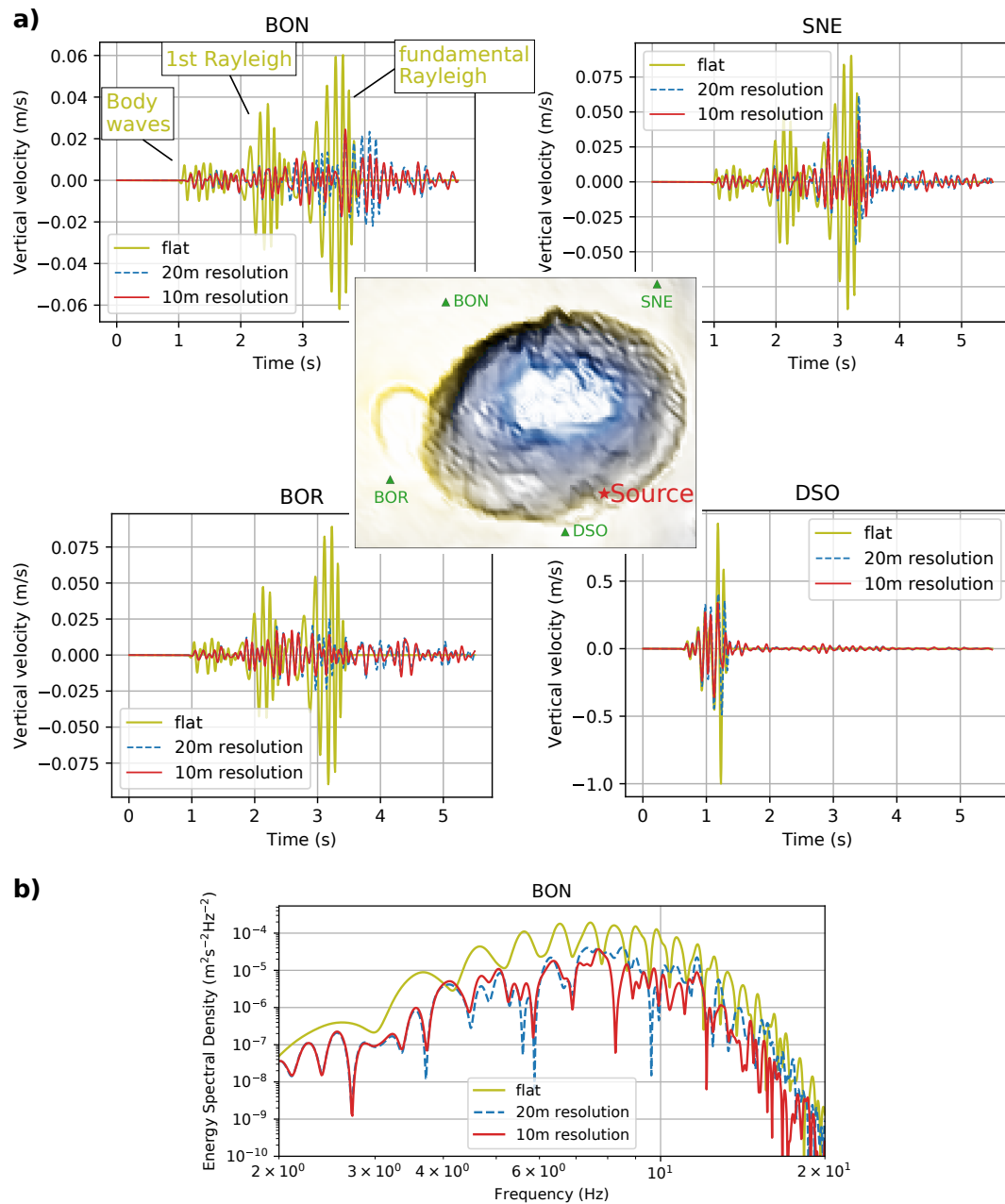


Figure 3. Influence of topography resolution on synthetic seismograms from the Lesage generic velocity model, recorded at stations BON, BOR, DSO and SNE. **a)** Comparison of synthetic seismograms (vertical velocity, normalized by maximum amplitude at closest station DSO) from the model with a flat surface, model with 20 m topography resolution (low-pass filtered with 30 m corner wavelength), and model with 10 m topography resolution. Seismograms recorded at stations BON, BOR, DSO, and SNE, surrounding Dolomieu crater. SEM configurations correspond to 7, 8, and 16 in Table A1. **b)** Corresponding spectra recorded at station BON.

275 model, wave packets corresponding to body waves, first-mode Rayleigh waves and fundamental-
 276 mode Rayleigh waves are well separated, but become less distinguishable when introduc-
 277 ing topography. Comparing the two models with topography, note that the first part of
 278 the wave-train is almost identical, which is related to body waves not being affected by

279 topography variations because of less interaction with the surface. Greater amplitude
 280 differences are found at later arrival times, suggesting that mainly slower surface waves
 281 of smaller wavelengths are affected by topography variations.

282 This assumption is supported by the spectra recorded at station BON, Figure 3b.
 283 Differences between the two models with topography become evident above roughly 5 Hz.
 284 This corresponds to a minimum wavelength of 116 m for the fundamental Rayleigh wave
 285 ($\lambda \approx 580 \text{ m s}^{-1} \div 5 \text{ Hz} \approx 116 \text{ m}$). If we conclude that wavelengths below 116 m are still
 286 sensitive to the change in topography resolution, then 1st-mode Rayleigh waves of above
 287 7 Hz are affected ($\lambda \approx 800 \text{ m s}^{-1} \div 7 \text{ Hz} \approx 114 \text{ m}$). This analysis suggests surface waves
 288 are sensitive to changes in topography resolution that are 5 times smaller than their wave-
 289 length.

290 The decrease in amplitude at all stations for the high-resolution topography (10 m)
 291 relative to the lower-resolution topography (20 m) suggests that more energy is scattered
 292 during the propagation of the surface waves and possibly lost within the subsurface. In-
 293 terestingly, S. J. Lee, Chan, et al. (2009) found the opposite when comparing waveforms
 294 on different topography resolutions for a source deep below the surface. This implies that
 295 the source position plays a major role in the effect of topography. On one hand, topog-
 296 raphy can increase ground shaking and thus trap energy close to the surface. On the other
 297 hand, in the case of waves traveling along the surface, the topography can increase scat-
 298 tering and thus prevent energy propagation. Similar conclusions were drawn by S. J. Lee,
 299 Komatitsch, et al. (2009), who investigated how topography effects are modulated by
 300 the source depth in regard to ground motion in a basin located behind a mountain range.

301 **3.4 Wave propagation from a vertical surface load**

302 To better understand wave propagation along the topography and the influence of
 303 the subsurface geology, snapshots of the wave field were examined. Simulations for all
 304 three velocity models were carried out on the domain with 20 m Dolomieu topography
 305 resolution and a vertical point force on the southern crater wall.

306 Figure 4 shows synthetic seismograms recorded on the surface along an array cross-
 307 ing the source position, Dolomieu crater and station BON (see inset for location of the
 308 array). Snapshots of the propagating seismic wave field on a cross-section along the ar-
 309 ray are shown below the seismograms. All amplitudes correspond to vertical ground ve-
 310 locity. In order to enhance visibility of the wave field over time, the simulations here are
 311 carried out without intrinsic attenuation.

312 For the simulation with the homogeneous domain (left column of Fig. 4), we can
 313 identify in the first snapshot at time $t = 0.8 \text{ s}$ the P-wave traveling downwards as be-
 314 ing the fastest wave with propagation direction parallel to the shown vertical ground ve-
 315 locity. At time $t = 1.6 \text{ s}$, the original S-wave is visible on the bottom of the cross-section.
 316 The S-wave can be identified because the direction of propagation is perpendicular to
 317 the vertical ground velocity. Just above, note the newly created S-wave (annotated as
 318 RS) that separated at the bottom of the crater from the Rayleigh wave because of the
 319 convex topography. Yet, part of the energy continues as a Rayleigh wave along the to-
 320 pography towards the rim of the crater. Also visible is a diffracted surface wave (anno-
 321 tated as Rd). It split from a wave front traveling towards station BOR and took a curved
 322 path along the flank of the crater. At time $t = 2.0 \text{ s}$ we can see this diffracted Rayleigh
 323 wave continuing outside the crater and arriving at station BON at a different azimuth
 324 than that of the Rayleigh wave that traveled diagonally across the crater and its rim (an-
 325 notated as Rf). The energy of Rayleigh wave Rf was partly reflected at the crater rim
 326 so that a new Rayleigh wave Rr traveled backwards through the crater. Up front (on the
 327 far right of the domain), a direct S-wave hits the surface and is partly reflected and con-
 328 verted to build a straight P-wave front traveling downwards at an oblique angle to the
 329 horizontal (annotated as SP).

330 Adding a low S-wave velocity layer (middle column in Fig. 4) drastically changes
 331 the wave field because of reflections within this layer and the dispersive character of Rayleigh

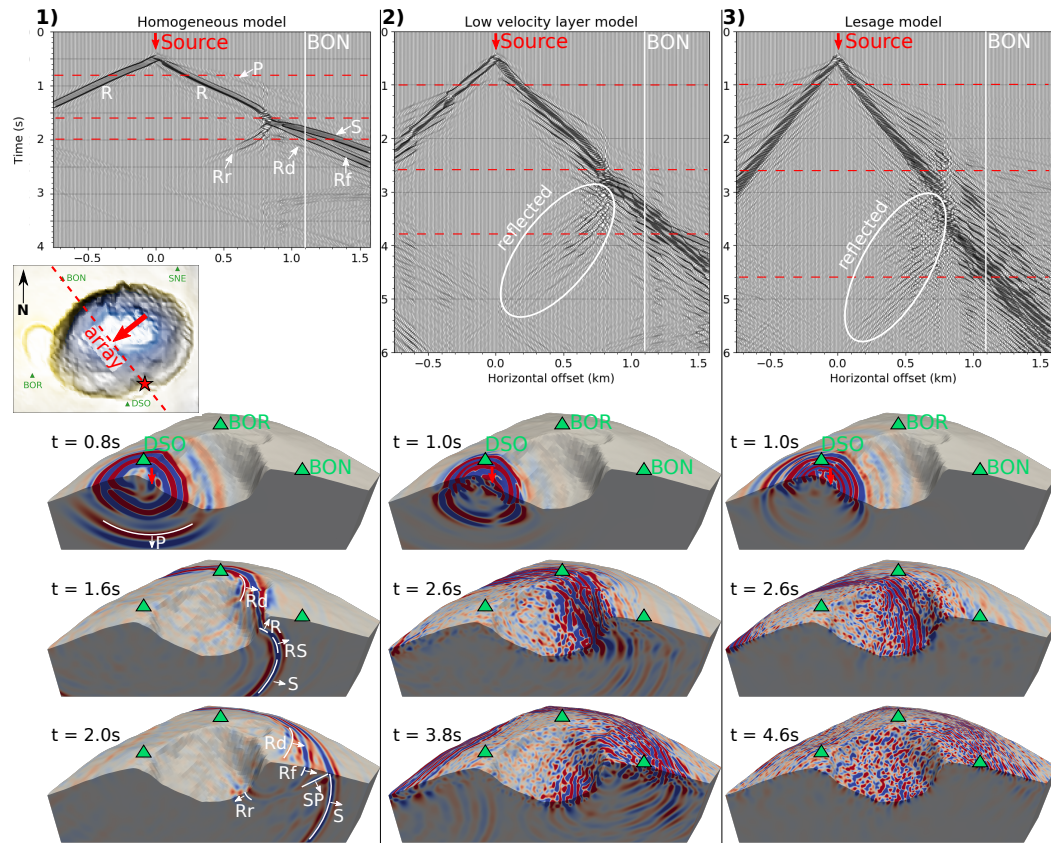


Figure 4. Wave propagation from a vertical surface load for different velocity models. Synthetic seismograms (*top row*) recorded at an array crossing the source, Dolomieu crater and station BON (see inset) for the **(1)** homogeneous model (*left*), **(2)** model with a shallow S-wave velocity layer (*middle*), and **(3)** Lesage generic velocity model (*right*). The seismic traces are normalized with respect to themselves and show vertical ground velocity. Snapshots of the wave field on cross-sections along the same array are shown below the seismograms, corresponding to the times marked by red dashed lines. Annotations denote P-wave (P), S-wave (S), P to S converted wave (PS), Rayleigh to S converted wave (RS), Rayleigh wave (R), reflected Rayleigh wave (Rr), diffracted Rayleigh wave (Rd), and diagonally traveled Rayleigh wave (Rf). SEM configurations correspond to 3, 6 and 11 in Table A1. The absence of intrinsic attenuation for enhanced visibility of wave propagation caused reflections from the boundary on the left at later times.

332 waves. Looking at the first 2.5 s of the synthetic seismograms, we observe a wave-train
 333 with a dispersive character overlaid by multiples. Compared to the homogeneous model,
 334 it is more complex and has a longer duration because of internal reflections within the
 335 low-velocity layer. At around $t = 2.6$ s, the waves hit the crater rim opposite the source
 336 and are partly reflected just like in the homogeneous case. The snapshots at times $t =$
 337 2.6 s and $t = 3.8$ s show, in contrast to the homogeneous case, a much more scattered
 338 wave field of irregular amplitude patterns. Similar to S. J. Lee, Chan, et al. (2009), who
 339 found characteristic patterns dependent on the resolution of the imposed topography,
 340 the characteristic length of these patterns is likely to be related to the resolution of the
 341 topography and the flat element surfaces with 20 m side lengths.

342 For the Lesage generic velocity model (right column of Fig. 4), the majority of energy
 343 stays close to the surface of the domain because of the velocity gradient. Scatter-

344 ing of the wave field over the topography is even greater than for the case with a low-
 345 velocity layer and the duration of shaking is increased. From synthetic seismograms (top
 346 right of Fig. 4), we can still identify the outward propagation of energy as well as the
 347 reflection of part of the energy at the crater rim. The analysis of the simulations shows
 348 that a single impact can produce a complex wave field caused by the surface topogra-
 349 phy and the underlying velocity model.

350 Regarding scattering from 3D medium heterogeneities, results from simulations in
 351 which the Lesage velocity model is randomly perturbed by heterogeneities with a stan-
 352 dard deviation of 10% (with maximum excess of 43%) show barely affected synthetic sig-
 353 nals, see Supporting Information, indicating that the effective medium is not changed
 354 significantly. This example does not prove that scattering is weak as it is possible to de-
 355 sign a distribution of characteristic correlation lengths that completely changes the wave-
 356 forms. However, studies at Dolomieu crater lead us to believe that this is not the case;
 357 e.g. Hibert et al. (2011) found that rockfall seismic signals do not exhibit a coda but that
 358 their duration corresponds to the rockfall propagation time on videos and Kuehnert et
 359 al. (2020) tracked rockfall trajectories using simulated inter-station energy ratios with
 360 the smooth Lesage velocity model.

361 4 Influence of topography on simulated wave propagation

362 Seismic amplitudes carry crucial information on the seismic source and can be used
 363 to infer source locations and acting forces. However, as can be concluded from the sim-
 364 ulated wave propagation above, topography together with the underlying geology can
 365 strongly influence ground motion. Consequently, the measured amplitudes have to be
 366 interpreted according to both source properties (including the resulting radiation pat-
 367 terns) and propagation effects. In the following, topography induced amplification is quan-
 368 tified for different velocity models and different source directions. This can be helpful
 369 to better interpret the spatial distribution of amplitudes and eventually account for am-
 370 plified signals.

371 4.1 Amplification for a vertical source

372 In order to quantify topographic ground motion amplification, simulations on a model
 373 with topography are compared to a reference model with a flat surface. The compari-
 374 son is performed for both vertical peak ground velocity PGV_z and total kinetic energy
 375 E . Quantifying PGV amplification is important when interpreting signal amplitudes. How-
 376 ever, it does not measure the increased complexity and duration of recorded waveforms
 377 caused by scattering and diffraction of the wave field along the topography. These ef-
 378 fects can be incorporated by calculating energy amplification. Also, frequency depen-
 379 dencies are not considered. For this reason we will later look at different frequency bands
 380 or determine spectral ratios when analyzing observed rockfall signals.

381 To quantify vertical PGV amplification, the maximum vertical ground velocity is
 382 measured at each point on the surface defined on a grid with 30 m spacing. The top row
 383 of Figure 5 shows the peak ground velocity ratios $PGV_{z,T}/PGV_{z,F}$ between the three
 384 velocity models with topography and the flat reference model.

385 Similarly, energy amplification is calculated at each grid point by the ratio E_T/E_F
 386 between the models with topography and the flat reference model, where E_i (with $i =$
 387 T, F) is a proxy of the total kinetic-energy density, defined as the square of the recorded
 388 ground velocity \vec{v} , integrated over the total signal duration d :

$$E_i = \int_d (v_{x,i}^2(t) + v_{y,i}^2(t) + v_{z,i}^2(t)) dt, \quad (2)$$

389 The resulting energy amplification is shown in the bottom row of Figure 5 for the three
 390 different velocity models.

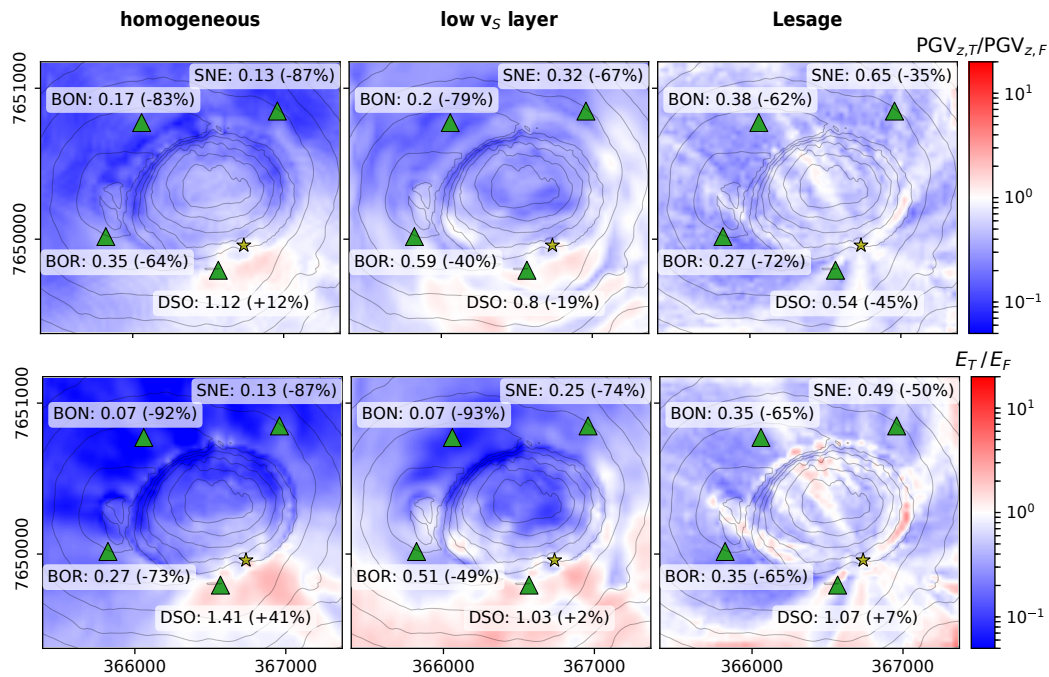


Figure 5. Topographic amplification from for a vertical point force. Amplification for of vertical PGV (*top*) and total kinetic energy (*bottom*) is calculated relative to a flat reference model for the homogeneous model (*left*), the model with a shallow low velocitylow-velocity layer (*middle*) and the Lesage generic velocity model (*right*). SEM configurations correspond to 1, 2; 4, 5; 7, 8 in Table A1. The yellow star illustrates the source position and green triangles mark station locations. Annotations give ratios measured at the station locations as well as the percentage of topographic amplification. Neighboring contour lines differ by 60 m in elevation.

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4.1.1 PGV amplification

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Analyzing the PGV amplification shown in the top row of Figure 5, the homogeneous model shows a contrast between source side of the crater and the opposite side: PGV is amplified on the source side and strongly deamplified on the far side. The amplification on the source side (+12% at DSO) can be explained by the simultaneous arrival of surface and direct waves emitted from the source. Deamplification on the far-side of the source (-83% at BON and -87% at SNE) can be interpreted as a shadow zone behind the crater related to the diversion of a major part of wave energy downwards into the subsurface because of the crater shape.

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For the model with the low-velocity layer, general amplification on the source side and deamplification on the far-source side of the crater are still present but less pronounced (deamplification at station SNE is reduced to -67%) and patterns become more complex (DSO is now deamplified by -19%). The introduction of a low-velocity layer causes more energy to stay at the surface and thus reduces the shadow zone behind the crater. The uneven topography together with the underlying low-velocity layer causes complicated reflections and wave conversions which lead to increased complexity of amplification patterns.

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The contrast between source side and far-source side of the crater decreases further for the Lesage generic velocity model (-45% at DSO, -62% at BON and -35% at SNE). As can be seen on the wave propagation snapshots in Figure 4, the gradient causes energy to stay close to the surface. Whereas a lot of energy is lost downwards because

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412 of the crater topography in the homogeneous model as well as in the low-velocity layer
 413 model, the velocity gradient in the Lesage generic model guides waves efficiently along
 414 the crater topography or back to the surface, which causes a more homogeneous ampli-
 415 fication pattern. Scattering away from the surface due to surface roughness as well as
 416 conversion from vertical to horizontal energy leads to an overall deamplification in ver-
 417 tical PGV. Still, because of focusing mechanisms of the 3D topography, ray-shaped zones
 418 of PGV amplification originating at the source can be observed.

419 **4.1.2 Energy amplification**

420 In general, the amplification patterns of kinetic energy (bottom row of Figure 5)
 421 show more contrast than the PGV ratios. This is because topography does not only in-
 422 fluence peak amplitude, but also the complexity and duration of the signal. For the ho-
 423 mogeneous model, amplification increases to +41% at DSO and decreases to -92% at
 424 BON. Behavior for the model with the low-velocity layer is very similar. For the Lesage
 425 generic model, the ray-shaped zones of amplification are considerably more pronounced
 426 than for the case of PGV amplification. Given that horizontal ground velocity is con-
 427 sidered when computing the kinetic energy, this observation suggests that topography
 428 guides both vertical and horizontal energy along the same paths. Note also the increased
 429 amplification at parts of the crater cliff ridge, possibly due to the discussed reflection of
 430 Rayleigh waves at these positions.

431 Changing the velocity model modifies the wavelengths, which presents an alterna-
 432 tive explanation for the observed differences in the amplification patterns. This expla-
 433 nation was discarded after verifying that the differences still remain for bandpass filtered
 434 results, comparing amplification patterns from the different velocity models for coincid-
 435 ing wavelengths as done in Appendix B.

436 **4.2 Amplification for a horizontal source**

437 Only vertical surface loads are considered above. However, the rockfall generated
 438 basal forces on the ground can also have horizontal components. Here we show ampli-
 439 fication patterns for a horizontal source using the Lesage generic velocity model. Fig-
 440 ure 6 illustrates vertical PGV amplification (*left*) and energy amplification (*right*) for
 441 a wave field generated by a horizontal surface force polarized in the north direction.

442 A strong directionality is visible in the PGV amplification pattern. This is because
 443 for the flat reference model, a horizontal source does not generate vertical seismic en-
 444 ergy perpendicular to its direction. Topography however can change this by conversion
 445 from transverse energy or diffraction of wave paths.

446 The directionality patterns are no longer visible when analyzing the amplification
 447 of total kinetic energy. This is because all components of the measured ground veloc-
 448 ity are considered in the energy calculation. The energy amplification pattern is simi-
 449 lar to the one for the vertical source as shown above in Figure 5. This suggests that to-
 450 pography guides seismic energy on trajectories along the surface that mainly depend on
 451 the source position rather than the source direction. We will further discuss this hypoth-
 452 esis later when studying inter-station spectral ratios of real rockfall signals.

453 **4.3 Surface roughness and crater geometry**

454 The amplification patterns observed in the previous section are characterized by
 455 complex spatial distributions. We will now perform tests on domains with synthesized
 456 surface topographies in order to better understand the contributions of certain geomet-
 457 ric features to the amplification pattern. More concretely, we will study a planar sur-
 458 face with natural roughness as well as synthetic crater shapes of different depths and cur-
 459 vatures. Surface roughness and crater dimensions are defined to resemble our study site
 460 on Piton de la Fournaise volcano. The initial domain is a cube of size 2360 m × 2360 m ×

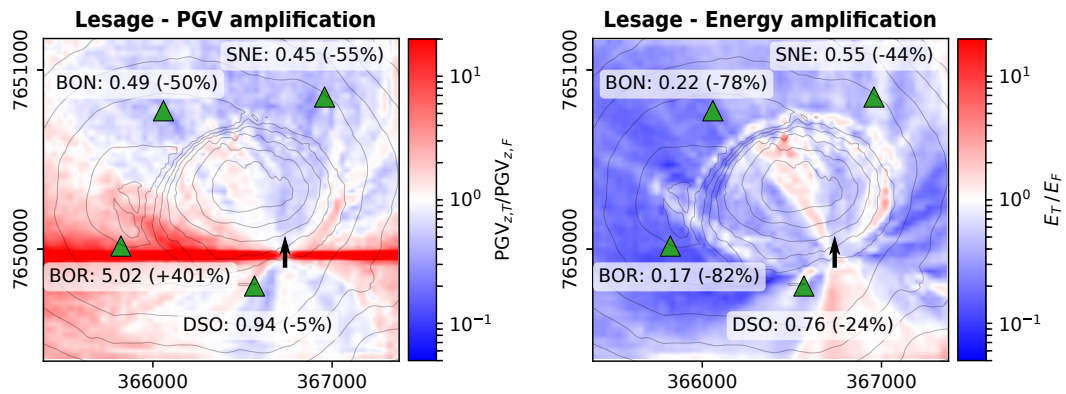


Figure 6. Topographic amplification for a horizontal point force in the north-direction. Amplification of vertical PGV (*left*) and total kinetic energy (*right*) is calculated relative to a flat reference model for the model with the Lesage generic velocity profile. SEM configurations correspond to 9 and 10 in Table A1. The black arrow illustrates the source position and its direction. Green triangles mark station locations. Annotations give ratios measured at the station locations as well as the percentage of topographic amplification. Neighboring contour lines differ by 60 m in elevation.

600 m, meshed by elements with 20 m side lengths. The subsurface medium of all domains corresponds to the Lesage generic velocity model. As above, a 7 Hz Ricker wavelet is used as the surface point force.

The domain with a planar rough surface is constructed from an area of the DEM at Piton de la Fournaise volcano and band-pass filtered at corner wavelengths of 40 m and 100 m. In this way, minimum and maximum wavelengths of the fundamental Rayleigh wave in the Lesage generic model are below and above the range of topography wavelengths, respectively (i.e. $\lambda_{15\text{ Hz}} \approx 390\text{ m s}^{-1} \div 15\text{ Hz} = 26\text{ m}$ and $\lambda_{5\text{ Hz}} \approx 580\text{ m s}^{-1} \div 5\text{ Hz} = 116\text{ m}$). To design a typical crater geometry, we use the equation proposed by Soontiens et al. (2013). However, using a smooth, symmetric crater shape results in symmetric interferences. To avoid artificial amplification patterns of perfect symmetry, the above defined surface roughness is added to the elevation values of the synthetic crater shape. The corresponding SEM meshes are shown in the Supplementary Information.

Figure 7a compares synthetic seismograms recorded along arrays on the domains with a flat surface, a planar rough surface, and crater topography.

For the model with the flat domain, we can identify dispersive fundamental and first-mode Rayleigh waves as well as body waves. Introducing surface roughness leads to strong scattering and hence prolonged ground shaking. The two Rayleigh modes are no longer clearly separated, even though the propagation of the main energy from the fundamental mode can be identified. Introducing the crater topography adds more complexity. In particular the steep crater walls distort the propagating wave field, as already observed for the real crater topography (see Fig. 4).

We now investigate the effect of surface topography at different frequency bands. For this, we quantify as before the amplification of total kinetic energy with respect to the flat reference model. Note that here we present energy instead of PGV as it accounts for both amplitudes and prolonged ground shaking and hence gives a more general picture. Figure 7b shows energy amplification on both the rough planar domain and the domain with a synthetic crater in the 3-7 Hz and 13-17 Hz frequency bands.

We can see that both these frequency bands are influenced by the rough planar surface. As already indicated, the rough topography is band-pass filtered at corner wavelengths 40 m and 100 m and the minimum and maximum wavelengths of fundamental Rayleigh waves are below and above the range of topography wavelengths, respectively.

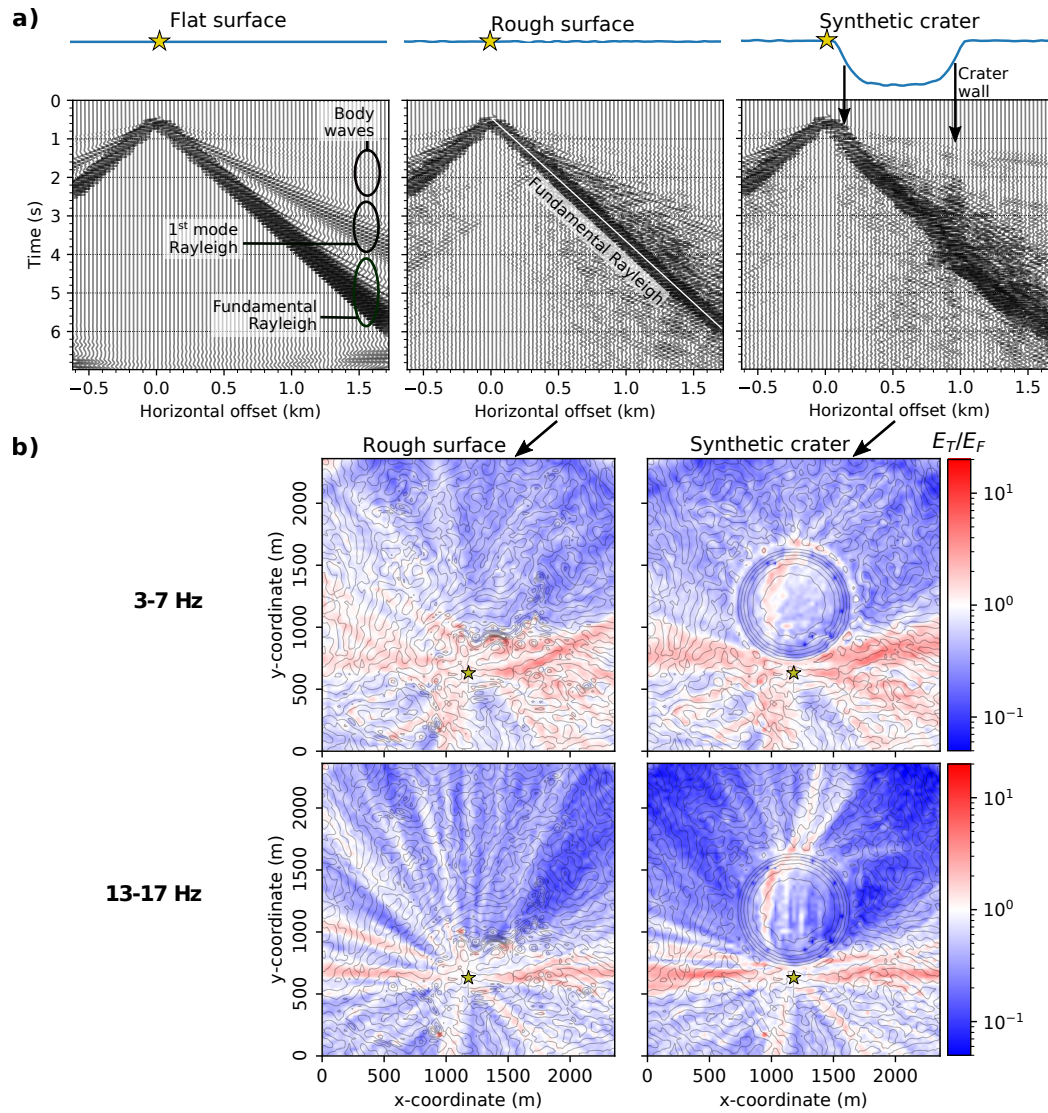


Figure 7. **a)** Synthetic seismograms of vertical ground velocity from models with a flat surface (*left*), a rough planar surface (*middle*), and a synthetic crater shape (*right*). Seismograms are normalized and recorded along the surface profiles illustrated by the blue curves on top. The yellow star marks the position of the vertical source. The spurious reverberations observed for the flat surface model after the signal (> 6 s) are trimmed for the analyses. **b)** Energy amplification in different frequency bands for the model with a rough surface (*left*) and with a synthetic crater (*right*) relative to the flat reference model for frequency bands 3-7 Hz and 13-17 Hz. The arc-like features at 3-7 Hz in the top corners are numerical artefacts. SEM configurations correspond to 12, 13 and 14 in Table A1.

493 We observe ray-shaped zones of amplification which are blurred in the lower-frequency
 494 band and become sharper towards higher frequencies, because of the shorter interfering
 495 wavelengths. The variations of topography seem to guide energy along these ray paths.
 496 In contrast, some areas of pronounced topography variation (visible by the densification
 497 of contour lines) seem to shield the propagation of energy and cause shadow zones be-
 498 hind them. This can for example be observed in the north-east direction of the source.

499 Analyzing the energy amplification on the domain with the synthetic crater, we re-
 500 cognize similarities with the amplification patterns of the previously analyzed planar rough
 501 surface. This is because the same surface roughness is used and its imprint is now su-
 502 perimposed on the amplification caused by the crater topography. Globally, the wave
 503 field is deamplified behind the crater (as seen from the source position). Higher frequen-
 504 cies seem to be more affected by this than lower frequencies. Nonetheless, even at high
 505 frequencies, paths of amplified energy can traverse the crater. This phenomenon seems
 506 to be caused by a coupled effect from small-scale (i.e. roughness) and large-scale (i.e. crater)
 507 topography variations and is similar to the amplified ray-paths observed on the Dolomieu
 508 crater topography (compare to Figure 5).

509 The sensitivity of the amplification pattern to variations in crater depth and cur-
 510 vature was studied. The parameters were chosen so that on one hand crater depth varied
 511 by ± 0.3 (from small to big) with fixed curvature and on the other hand maximum
 512 curvature varied by ± 0.3 (from weak to strong) with fixed crater depth. The resulting
 513 profiles and their curvatures are compared to a profile through Dolomieu crater on the
 left-hand side of Figure 8.

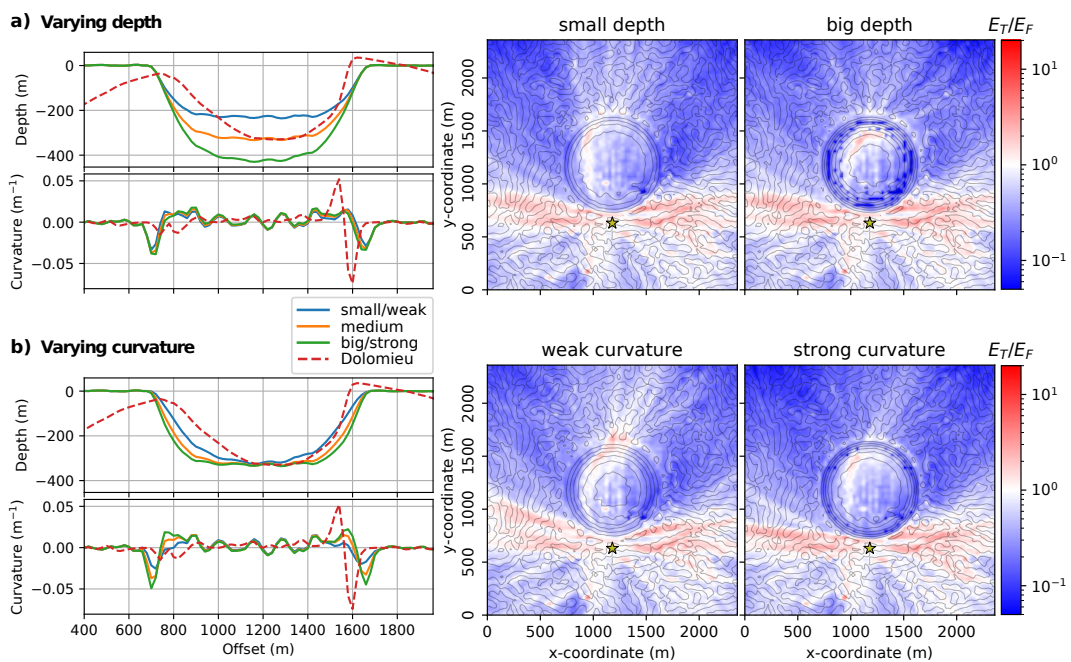


Figure 8. *Left:* Profiles through the synthetic crater topographies. **a)** Crater depths vary by ± 0.3 from small to big with fixed curvature. **b)** Curvatures vary by ± 0.3 from weak to strong with fixed crater depth. Red dashed lines correspond to a profile through Dolomieu crater and its corresponding curvature. *Right:* Comparison between energy amplification for crater geometries with smallest and biggest depths as well as weakest and strongest curvature. Contour lines mark elevation differences of 50 m and the yellow star denotes the source. Note that spurious blue dots inside the crater (especially at steep flanks for the big depth) were caused by numerical measurement problems at these positions. SEM configurations correspond to 12 and 14 in Table A1.

514 The energy ratios from the simulations on the domains with synthetic crater shapes
 515 are shown on the right-hand side of Figure 8 for the whole frequency range. The ampli-
 516 fication pattern varies slightly going from small depth to big depth (Figure 8a). The biggest
 517 change is observed behind the crater directly opposite the source. Amplification decreases
 518

519 at this point with increasing crater depth. In contrast, increased amplification is observed
 520 inside the crater. These changes in the amplification patterns might be related to inter-
 521 ference caused by the symmetric crater form. Going from weak to strong curvature (Fig-
 522 ure 8b), the shadow zone behind the crater increases more strongly. This is not only true
 523 directly opposite the source position but also diagonally across the crater, suggesting that
 524 the increased crater curvature shields off more energy by reflecting or deflecting the wave
 525 field sideways or downwards into the subsurface.

526 The analyses suggest that variations of curvature have stronger effects on ground
 527 motion than variations of crater depth for models tested. It is important to note that
 528 the wave field is influenced by topography features of scales both below and above the
 529 seismic wavelength. This was observed for the experiments with a planar rough surface
 530 as well as with the synthetic crater with dimensions (~ 800 m diameter, ~ 300 m depth)
 531 largely exceeding the seismic wavelengths.

532 The experiments on the synthetic model surfaces explored effects of individual as-
 533 pects of the topography on the wave field. The insights acquired can be transferred to
 534 our study site at Piton de la Fournaise volcano as similar scales were chosen deliberately.
 535 Having said this, note that the overall effect on the wave field is governed by the whole
 536 configuration and cannot be reduced to an individual feature (e.g. only small-scale ver-
 537 sus only big-scale topographic variations). At the same time, the relative position of the
 538 source and receiver plays a defining role, i.e. topography related amplitude modifications
 539 at a given station can only be predicted if the source position is known.

540 **5 Seismic signals from rockfalls at Dolomieu crater**

541 We will now study observed seismic signals generated by rockfalls at Dolomieu crater.
 542 As the influence of the topography changes with the source position, we analyze the sig-
 543 nals at specific times corresponding to specific rockfall positions. First we will investi-
 544 gate spectral ratios between stations of time windowed rockfall signals. The objective
 545 is to clarify as to whether simulations can reproduce the observed spectral ratios when
 546 taking into account topography. Subsequently we will focus on a single block impact,
 547 identifying its seismic signature and comparing observed and simulated amplitudes by
 548 estimating the generated impact force using Hertz contact theory.

549 **5.1 Observed spectral ratios between stations**

550 For this analysis, we select three rockfalls with similar trajectories on the south-
 551 ern crater wall corresponding to rockfall location **1** in Figure 1b. The trajectories of the
 552 rockfalls were identified from camera recordings. Snapshots of the three events are shown
 553 in Figure 9 together with an image of the whole trajectory reconstructed from differences
 554 of successive snapshots. Below, the corresponding seismic signals recorded at stations
 555 BON, BOR, DSO and SNE are presented.

556 Station DSO shows the strongest amplitudes, especially in the beginning of the rock-
 557 fall. This is because the three rockfalls start very close to this station. BON contains
 558 the smallest amplitudes, being the furthest station and on the opposite side of the crater.
 559 The dynamics of the three events are not entirely identical. Event 1 consists of a single
 560 boulder bouncing down towards the bottom of the crater while other blocks follow with
 561 a time lag of around 15 s. In contrast, event 2 consists of two blocks closely following each
 562 other down with a time lag of only 4 s, as can be seen on snapshot 2b. Event 3 consists
 563 of a main boulder with a smaller block following much later with a lag of about 50 s.

564 Despite these differences, we compare spectral ratios between stations in time win-
 565 dows R1, R2 and R3 during which the main blocks moved within identical areas. The
 566 spectral ratios are computed from the measurements at stations BOR, DSO (vertical com-
 567 ponent only) and SNE with respect to station BON (note that BON is selected as the
 568 reference station as it turns out it is the least affected by local site effects). In order to
 569 avoid spurious fluctuations, the spectra are smoothed as proposed by Konno and Ohmachi

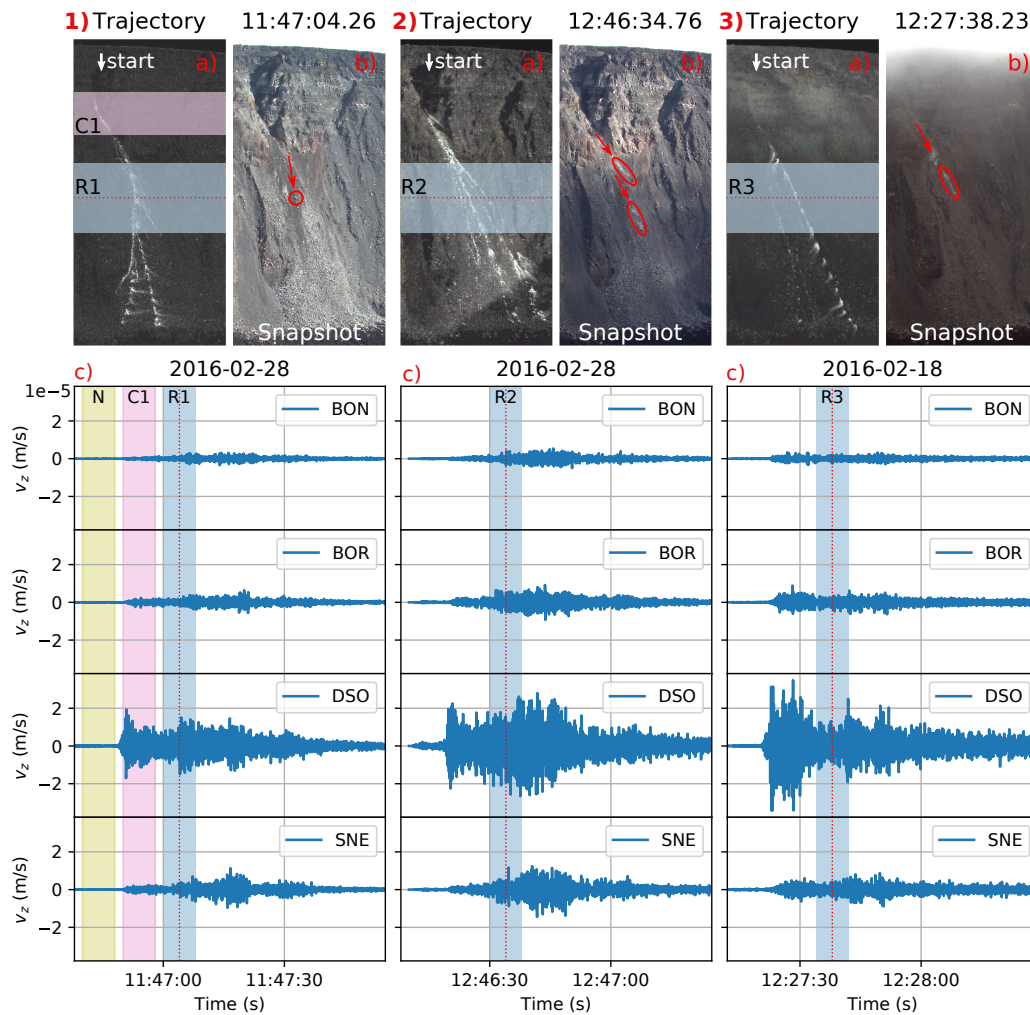


Figure 9. Three similar rockfalls on the southern wall of Dolomieu crater, corresponding to rockfall location 1 in Figure 1b. The events occurred on: **1)** February 28, 2016 at around 11:47, **2)** February 28, 2016 at around 12:46, and **3)** February 18, 2016 at around 12:27. **a)** The total trajectory of each event (seen from camera SFRC). The approximate starting positions at the top of the crater wall are indicated by white arrows. **b)** Snapshots (seen from camera SFRC) at a chosen time at which all three rockfalls are at comparable positions. **c)** The rockfall seismic signals, the red dotted lines indicating the times of the snapshots. Time windows R1, R2, and R3 (blue-shaded zones) are defined ± 4 s around these times. The corresponding locations of these time windows are also indicated as blue-shaded zones on the trajectories. The same holds for reference time window C1 (magenta-shaded zone), which corresponds to the beginning of event 1. Noise time window N is taken from recordings before event 1.

570 (1998) before calculating the ratios. The obtained curves are shown as dark blue lines
 571 (TW-R1, -R2, -R3) in Figure 10 for vertical- (*top*), north- (*middle*) and east- (*bottom*)
 572 components. Note the similar behavior of the spectral ratios for each of the events and
 573 for each component.

574 Comparison to spectral ratios from noise recordings (TW-N), from the beginning
 575 of event 1 (TW-C1), and from a rockfall that occurred at a different position in the crater
 576 (TW-C2, corresponding to trajectory 2 in Figure 1b), shows partly strong deviations from

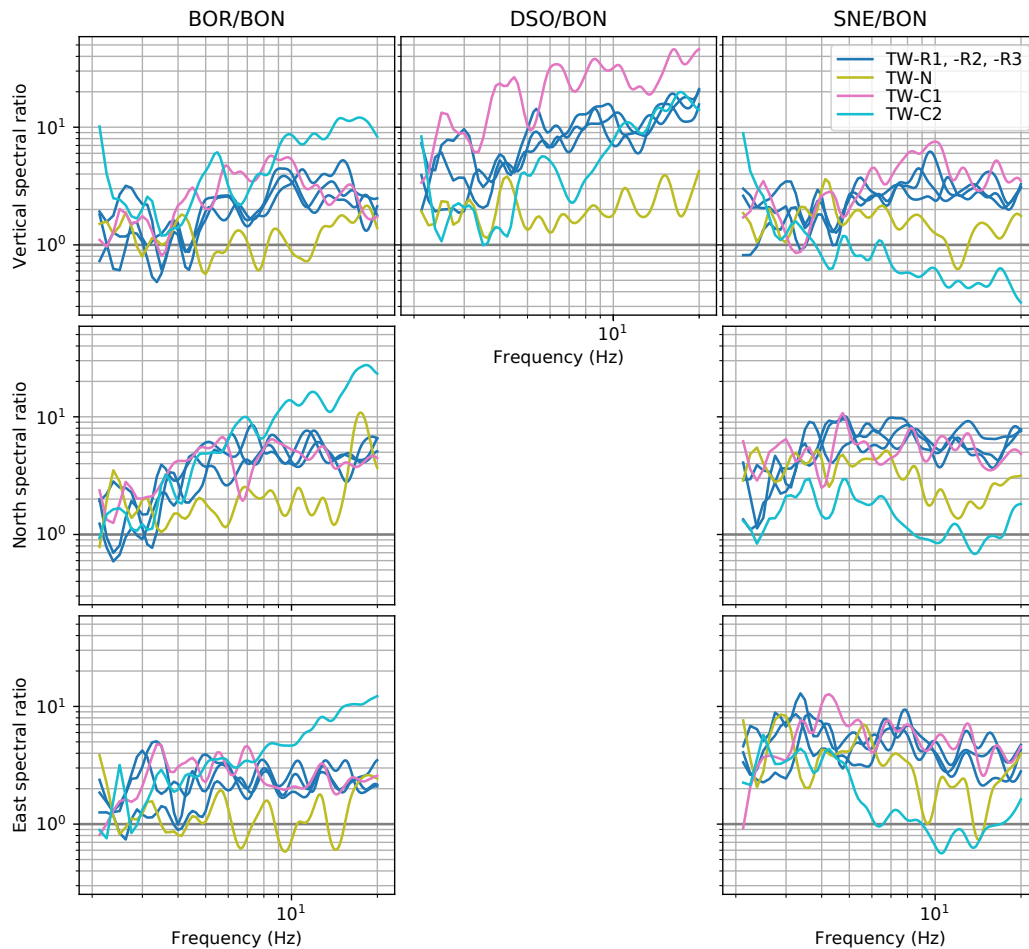


Figure 10. Spectral ratios from rockfall seismic signals recorded at stations BOR (3 components), DSO (1 component) and SNE (3 components) relative to station BON for vertical- (*top*), north- (*middle*) and east- (*bottom*) components. Time windows TW-R1, -R2, and -R3 correspond to rockfalls 1, 2, and 3 as defined in Figure 9. Time windows TW-N and TW-C1 correspond to noise recordings and the beginning of rockfall 1, respectively. Time window TW-C2 is taken from a rockfall on the southwestern crater wall, corresponding to rockfall location **2** in Figure 1b.

577 curves R1, R2 and R3. This provides evidence that the spectral ratios are indeed char-
 578 acteristic of the position of the rockfall seismic source. The same analysis is carried out
 579 in Appendix D for rockfalls in the southwest, leading to the same conclusion.

580 **5.2 Comparison of observed and simulated spectral ratios**

581 The seismic source of a rockfall can be very complex as multiple impacts of differ-
 582 ent magnitude can occur simultaneously at different positions. Hence, it is very difficult
 583 to correctly simulate the rockfall seismic signal, especially at high frequencies. For this
 584 reason, spectral ratios between stations are very convenient to compare real and synthetic
 585 signals. In this way, the signature of the source is removed from the signal and solely prop-
 586 agation path effects remain. Nevertheless, we have to keep in mind two points when compar-
 587 ing observed and simulated spectral ratios.

588 Firstly, local subsurface heterogeneities can modify recorded amplitudes and thus
 589 influence inter-station ratios. These geological site effects are not considered in the sim-

590 uations. Therefore, in order to enhance comparability between the observed and sim-
 591 ulated spectral ratios, the recorded signals are corrected using site amplification factors
 592 estimated from volcano-tectonic (VT) signals. The spectral amplification curves are cal-
 593 culated and discussed in Appendix C where we also show a comparison between simu-
 594 lated and uncorrected observed spectral ratios. The observed inter-station ratios presented
 595 here are corrected by deconvolution of the recorded signals with the corresponding am-
 596 plification factors.

597 Secondly, different source directions cause different radiation patterns. This is il-
 598 lustrated in Figure 11a, where a force on a flat surface is polarized in vertical (top) and
 599 horizontal (bottom) directions. If the radiation pattern is not radially symmetric, which
 600 is only the case for vertical ground motion from a vertical source, the spectral ratios are
 601 affected depending on the azimuthal position of the respective receivers. The direction
 602 of a rockfall seismic source depends both on the rockfall dynamics and on the underly-
 603 ing slope. The generated forces from a boulder impact are schematically illustrated in
 604 Figure 11b. The resulting force F_r is composed of a force F_n normal to the slope and
 605 a force F_t tangential to the slope, which depends on the slope angle, the direction of move-
 606 ment and the friction between the moving mass and the ground.

607 In order to analyze the influence of the source direction on the spectral ratios, we
 608 compare a vertical force to a normal force and a tangential force. Note that we assume
 609 that the tangential force is parallel to the slope of steepest descent. To consider a spa-
 610 tially distributed source in the simulations, the mean spectral ratio is calculated from
 611 a selection of multiple sources. This makes it possible to simultaneously evaluate the sen-
 612 sitivity of the curves to the source positions. Seven source positions are picked from a
 613 grid with 10 m spacing (see Fig. 1b, *picked sources*). The area corresponds to the region
 614 in which rockfalls 1, 2, and 3 are present during time windows R1, R2, and R3, respec-
 615 tively (see Fig. 9).

616 **5.2.1 Simulated spectral ratios for a model with a flat surface**

617 Figure 11c compares spectral ratios BOR/BON from the observed signals with sim-
 618 ulated ratios of differently polarized sources on a model with a flat surface. The source
 619 directions are determined from the slope of Dolomieu topography at the corresponding
 620 position before implementation on the flat domain. For spectral ratios of the vertical com-
 621 ponent (*left*), a tangential force direction results in much smaller values compared to the
 622 other sources. As the slope dips northwards, the tangential force is orientated in the north-
 623 direction. Station BOR is located west of the source position, which is transverse to the
 624 source direction. For this reason, a smaller signal amplitude is measured at station BOR
 625 in comparison with station BON (ratio < 1), even though BOR is slightly closer to the
 626 source. Nevertheless, the tangential force also contains a vertical component that ensures
 627 that the ratio is of the same magnitude as the observed ratios. This is different for the
 628 spectral ratios of the north component (*middle*), where a striking discrepancy of more
 629 than one order of magnitude results from the vertical source. A vertical force does not
 630 generate horizontal transverse energy which is why almost no signal is recorded on the
 631 north component at station BOR located eastwards. For the east spectral ratios (*right*),
 632 the tangential force again shows the strongest deviation for reasons similar to those for
 633 the vertical component spectral ratios.

634 **5.2.2 Simulated spectral ratios for a model with topography**

635 As opposed to the model with a flat surface, a model with the Dolomieu crater to-
 636 pography results in a good agreement between simulated and observed spectral ratios
 637 (see Figure 12). Furthermore, very similar values can be observed when comparing the
 638 simulations with different source directions, especially towards higher frequencies. This
 639 indicates that the spectral ratios are in this case not dominated by the direction of the

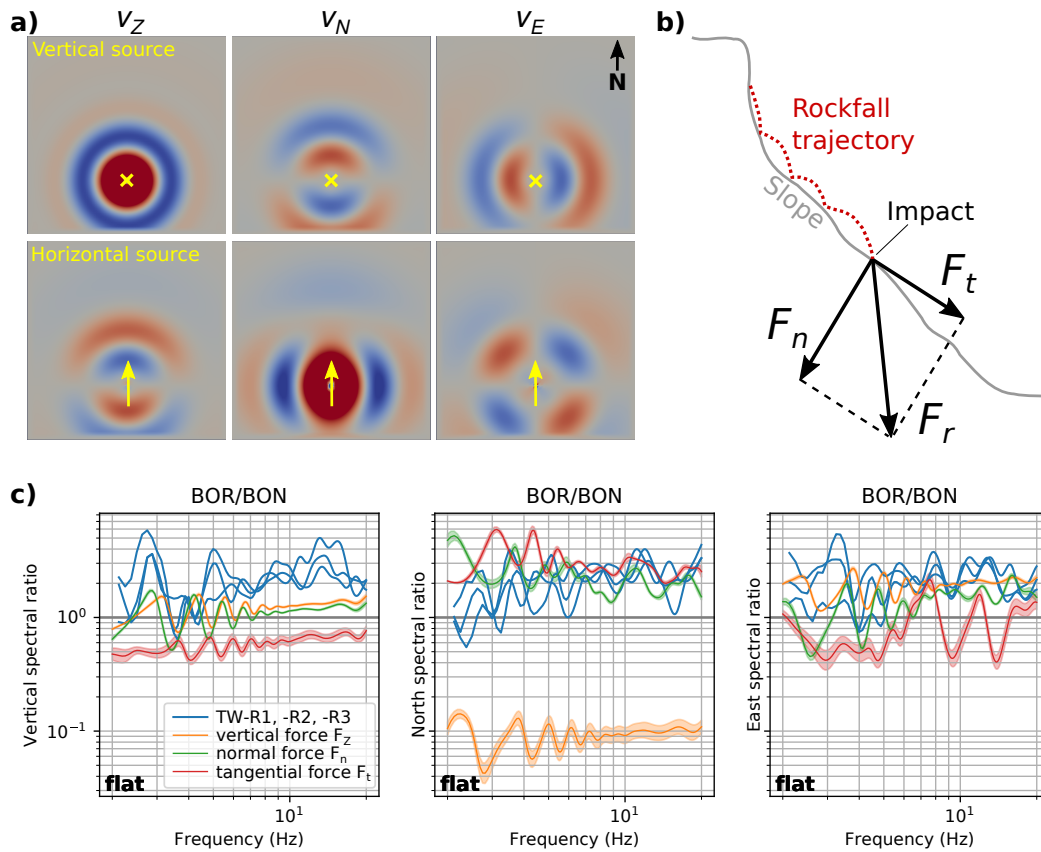


Figure 11. **a)** Seismic radiation patterns from a vertical source (*top*) and from a horizontal force (*bottom*) for ground velocity v_Z , v_N , and v_E of vertical-, north-, and east-component, respectively (red for positive and blue for negative amplitudes). **b)** Forces generated by a rockfall impact. The red dotted line illustrates the trajectory of a bouncing boulder. The impact generates force F_n normal to the slope. Depending on the boulder velocity tangential to the slope and on the friction coefficient μ , a tangential force $F_t = \mu F_n$ is generated (assuming Coulomb friction). Normal and tangential forces add up to resulting force F_r . **c)** Comparison of spectral ratios BOR/BON from real signals TW-R1, -R2, -R3 (as in Fig. 10, site-effect corrected) and from simulations on the flat domain with varying source direction: vertical force, normal force and tangential force according to the Dolomieu topography at the corresponding position. The shaded zones of the simulated ratios indicate the standard deviation around the mean value from seven neighbouring source positions (Fig. 1b, *picked sources*). SEM configurations correspond to 15 in Table A1.

640 source (and the corresponding produced radiation pattern) but rather by propagation
 641 along the topography.

642 Greater deviations between the different source directions are found at lower frequ-
 643 quencies, such as for example on the north component of ratio BOR/BON below 3 Hz.
 644 Assuming fundamental Rayleigh waves, this corresponds to wavelengths above 250 m ($\lambda \approx$
 645 $750 \text{ m s}^{-1} \div 3 \text{ Hz}$). With a distance of around 500 m between the source position and
 646 station BOR, it is likely that these low-frequency waves have not traveled enough wave-
 647 lengths in order to be completely dominated by propagation along the topography.

648 Analyzing the sensitivity of the ratios to the source position, generally larger stand-
 649 dard deviations (shaded zone of uncertainty around the mean) are present after intro-

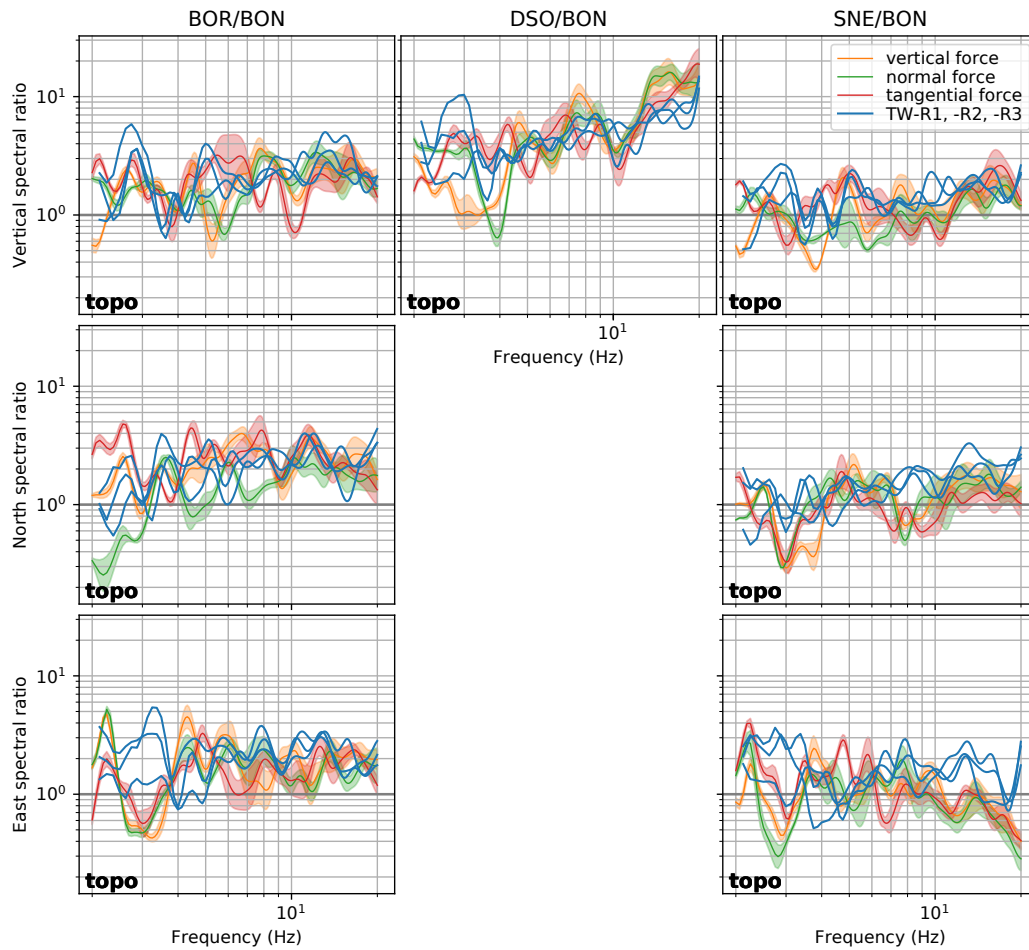


Figure 12. Spectral ratios BOR/BON, DSO/BON and SNE/BON calculated from real signals TW-R1, -R2, -R3 (as in Fig. 10, site-effect corrected) and from simulations on the domain with Dolomieu topography for the vertical (*top*), the north (*middle*), and the east (*bottom*) component. Simulations are carried out with varying source directions: vertical force, force normal to the slope and force tangential to the slope. The shaded zones of the simulated ratios indicate the standard deviation around the mean value from seven neighbouring source positions (Fig. 1b, *picked sources*). SEM configurations correspond to 16 in Table A1.

650 ducing topography compared to the results for the flat model in Figure 11c. This means
 651 that a slight change of source position allows more variability of the ratios when consid-
 652 ering topography and can eventually better explain the observed spectral ratios.

653 Clearly, the spectral ratios also depend on the relative source-receiver distance. For
 654 example, the high values of ratio DSO/BON result from the fact that the source is very
 655 close to station DSO. Furthermore, the values increase towards higher frequencies. This
 656 is related to the attenuating properties of the medium that cause the amplitudes of higher
 657 frequencies to decrease faster with the distance traveled than lower frequencies.

658 The analysis suggests that the spectral ratios are characteristic of the source posi-
 659 tion and dominated by propagation along the topography rather than by the radiation
 660 patterns caused by the source directions. To further validate this hypothesis, the same
 661 comparison between observations and simulations is carried out in Appendix D for rock-
 662 falls located on the southwestern crater wall.

663

5.3 Seismic signature of a rockfall impact

664

We will now analyze in detail the seismic signal generated by the single impacts of a rockfall at Dolomieu crater. Interpretation of the signal characteristics is based on comparison with synthetic signals simulated on models with and without topography. The comparison between observed and simulated signals has to be carried out very carefully because of the uncertainties of the seismic source and the propagation medium. It is important to emphasize that we do not want to reproduce the recorded signal but rather understand some of its features, such as for example arrival times, waveform complexity, and amplitudes.

672

For the analysis, a single boulder rockfall is chosen with well separated impacts that can be tracked on video. These criteria are fulfilled by an event that occurred on January 22, 2017, located on the northern crater wall. Figure 13 shows a camera snapshot of the rockfall at the time of impact N2 as well as the impact locations and the rockfall seismic signal recorded for the vertical component at the closest station BON. Two boul-

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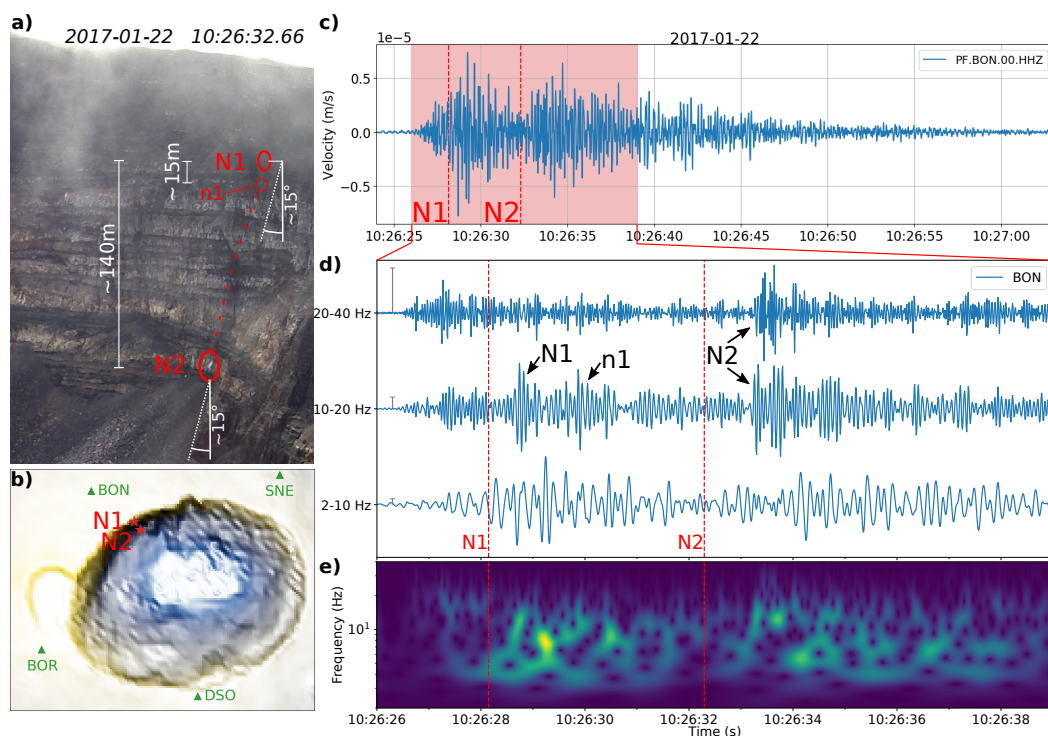


Figure 13. Single boulder rockfall on January 22, 2017. **a)** Camera snapshot taken shortly after impact N2. Estimated vertical distance between impacts and estimated slope angle to the vertical at impact positions. **b)** Location of impacts N1 and N2 in Dolomieu crater. **c)** Vertical ground velocity recorded at closest station BON in frequency band 2-40 Hz. The red-shaded area illustrates the time window of the graph below. Dashed lines mark impact times N1 and N2 estimated from the video. **d)** Comparison of frequency bands 2-10 Hz, 10-20 Hz, and 20-40 Hz. Signals are normalized to their maximum and the gray bars on the left indicate the relative scaling. **e)** Time-frequency representation of the rockfall signal (calculated using the Stockwell transform).

676

der impacts, N1 and N2, around 4s apart, are analyzed. A minor impact n1 is observed 1s after impact N1. It will be used later to estimate the fall velocity of the boulder. Note that the impact times are estimated from the video according to the appearance of small

677

678

679

680 dust clouds caused by the impacts. The time delay between the actual impact and the
 681 visibility of the dust cloud influences the accuracy of the impact time to a similar order
 682 of magnitude as the sampling time of 0.5 s between successive snapshots.

683 The broadband seismic signal of the rockfall shown in Figure 13c is characterized
 684 by two main lobes. These lobes are separated by a gap of low seismic energy at around
 685 10:26:32. During this gap, no impact is observed on the video. Thus, the boulder is in
 686 free fall before hitting the ground at impact location N2. Afterwards, the rockfall splits
 687 into several blocks that continue to move downwards on the debris cone of former rock-
 688 falls. At these later times it is very difficult to identify single impacts.

689 To better distinguish single impacts, the seismic signal is filtered in different fre-
 690 quency bands. Figure 13d compares signals band-pass filtered at 2-10 Hz, 10-20 Hz, and
 691 20-40 Hz. The relative scales of the normalized signals can be inferred from the gray bars
 692 plotted at the beginning of the signal as well as from the spectrogram below.

693 The signal filtered in the low frequency band (2-10 Hz) exhibits a smooth ampli-
 694 tude envelope. The two main lobes discussed above can be observed whereas no single
 695 pulses can be identified. This signal contains the strongest amplitudes and thus dom-
 696 inates the broadband signal. Short signal pulses emerge in the high-frequency bands. It
 697 is clear that seismic sources were already active before impact N1. Unfortunately, they
 698 could not be detected on the video. Impacts are possibly hidden behind the clouds on
 699 the top of the crater wall. A clear seismic pulse in the frequency range 10-20 Hz can be
 700 ascribed to impact N1. It arrives at the station around 0.5 s after the time determined
 701 from the video. A second pulse around 1 s later can be ascribed to impact n1. It con-
 702 tains slightly smaller amplitudes. The highest frequency band does not show signals clearly
 703 corresponding to these two impacts. This is different for impact N2. Both high-frequency
 704 bands show abrupt signal onsets around 1 s after the detection of impact N2 on the video.
 705 The following signal cannot be described as a single pulse but contains several peaks.
 706 This raises the question as to whether the source is made up of several impacts or if these
 707 peaks result from seismic wave propagation.

708 Another interesting observation concerns the impact-generated frequencies. As we
 709 can see, impact N1 is barely detectable in the highest frequency range (20-40 Hz), whereas
 710 impact N2 produces clear signals in both high-frequency bands (10-20 Hz and 20-40 Hz).
 711 Considering the changing source-receiver distance, we would expect the contrary as N1
 712 is slightly closer to station BON than N2. If we assume that the properties of the boul-
 713 der and of the underlying ground are identical for both impacts, the change in frequency
 714 content must be related to the impact velocity. As the boulder accelerates between im-
 715 pact N1 and impact N2, the higher velocity at impact N2 results in a shorter collision
 716 time and therefore generates higher frequencies according to Hertz contact theory, in-
 717 troduced below.

718 **5.3.1 Hertz contact theory**

719 To predict relative amplitudes of signals generated by impacts N1 and N2, the re-
 720 spective impact forces of the boulder on the ground are estimated. Farin et al. (2015)
 721 used the theory of Hertz (1878) to describe the force of an elastic sphere impacting a solid
 722 elastic surface. After successfully applying the theory on seismic signals generated in lab-
 723 oratory experiments, they analyzed real-size rockfall experiments carried out by Dewez
 724 et al. (2010). Here we estimate the impact forces in a similar fashion, assuming a spher-
 725 ical boulder of radius R and mass m , where $m = \rho \frac{4}{3} \pi R^3$ with rock density ρ . The
 726 maximum impact force F_0 exerted by the sphere perpendicularly to the plane can then
 727 be expressed as (Johnson, 1989)

$$F_0 = \frac{4}{3} ER^{1/2} \delta_{\max}^{3/2}, \quad (3)$$

728 where δ_{\max} is the maximum indentation depth

$$\delta_{\max} = \left(\frac{15mv_n^2}{16ER^{1/2}} \right)^{2/5}, \quad (4)$$

729 with impact speed v_n normal to the plane. E is the effective Young's modulus $1/E^* =$
 730 $(1 - \nu_s^2)/E_s + (1 - \nu_p^2)/E_p$, where $\nu_s, \nu_p, E_s,$ and E_p are Poisson's ratio and Young's
 731 modulus of a sphere and an impacted plane, respectively.

732 Concerning the frequency content of the impacts, we analyze the contact duration
 733 of the impacts. As proposed by Johnson (1989), the temporal evolution of the Hertzian
 734 impact force F_H can be approximated by

$$F_H(t) \approx F_0 \sin(\pi t/T_c)^{3/2}, \quad 0 \leq t \leq T_c. \quad (5)$$

735 The force-time function and its frequency spectrum are shown in Figure 14 as a func-
 tion of impact duration T_c .

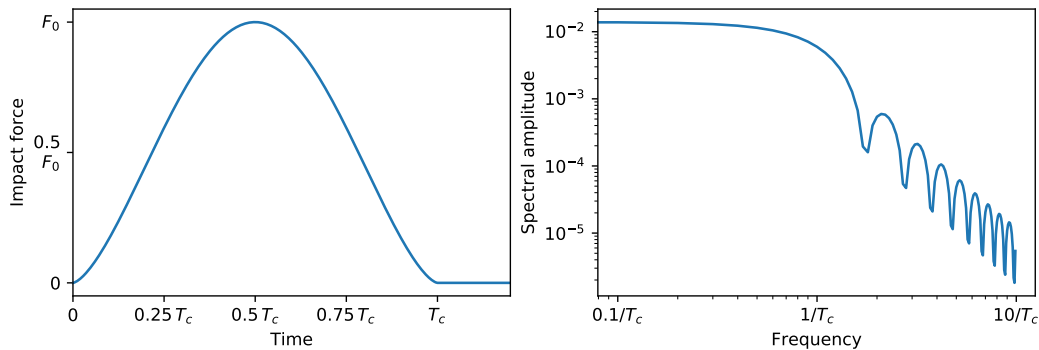


Figure 14. Hertzian impact force and corresponding frequency spectrum. *Left:* Hertzian force-time function F_H normalized by maximum impact force F_0 versus impact duration T_c , which represents the time during which the two bodies are in contact. *Right:* Frequency spectrum of the force-time function. The inverse impact time $1/T_c$ is related to the corner frequency f_c after which the spectral amplitude decays in the form of a power law.

736 The spectral amplitude decays in the form of a power law above corner frequency
 737 $f_c = 1/T_c$. Johnson (1989) showed that the impact duration can be approximated by
 738 means of maximum indentation depth δ_{\max} and impact normal speed v_n as
 739

$$T_c \approx 2.94 \frac{\delta_{\max}}{v_n}. \quad (6)$$

740 The values of impact parameters for N1 and N2 are estimated assuming that boulder
 741 and ground properties are identical between the two impacts. After careful consid-
 742 eration, the uncertainties for all parameters is uniformly set to $\pm 50\%$, which is found to
 743 be reasonable as a maximum threshold from a physical perspective and also allows nu-
 744 merical comparison of the respective contribution of each parameter on the total uncer-
 745 tainty of impact force and corner frequency.

746 To estimate impact speed v_n normal to the slope, a sub-vertical fall of the boulder
 747 before collision is assumed for both N1 and N2 with vertical speed v_c at the time of
 748 collision. Slope angles at the impact positions are inferred from the DEM to be around
 749 $\alpha = 15^\circ$. The normal impact speed can then be calculated as $v_n = v_c \sin \alpha$. To esti-
 750 mate v_c for N1 and N2, height differences between the impacts are determined from the
 751 DEM using the impact positions estimated from the video. As labeled in Figure 13a, we
 752 find a height difference of around $H_1 = 15$ m between N1 and n1 and a height differ-
 753 ence of around $H_2 = 140$ m between N1 and N2. Impacts N1 and n1 are detected 1 s
 754 apart. Assuming an approximately constant velocity during this short time window, the
 755 vertical speed for impact N1 is $v_{c,1} = 15 \text{ m s}^{-1}$. For impact N2, acceleration during the
 756 long free fall cannot be neglected. The speed is thus derived by $v_{c,2} = v_{c,1} + (2g(H_2 -$

757 $H_1))^{0.5}$, where $g = 10 \text{ m s}^{-2}$ is acceleration due to gravity. Hence, a vertical speed $v_{c,2} =$
 758 65 m s^{-1} is found for impact N2. This leads to normal impact speeds of $v_{n,1} = 4 \text{ m s}^{-1}$
 759 and $v_{n,1} = 17 \text{ m s}^{-1}$ for impact N1 and N2, respectively.

760 The boulder size is approximated from camera snapshots. The dust cloud caused
 761 by impact N2 in Figure 13a has an estimated length of 5 m. As only the dust clouds and
 762 not the boulder itself can be seen on the video, the boulder size is assumed to be less than
 763 1 m. As a lower bound, a size bigger than 0.2 m is assumed necessary to generate a seis-
 764 mic signal clearly above the ambient noise level. We therefore estimate the boulder size
 765 to be the mid-point between these two limits, i.e. 0.6 m. As a perfect sphere is consid-
 766 ered in the calculations, the effective radius is hence estimated to be $R = 0.3 \text{ m}$.

767 The rock density is estimated to be $\rho = 2000 \text{ kg m}^{-3}$, as in the simulations, which
 768 results in a boulder mass of $m = 226 \text{ kg}$. A typical effective Young's modulus of $E =$
 769 10 MPa is applied as proposed by Farin et al. (2015).

770 The maximum impact force F_0 can now be calculated using equation 3. We find
 771 84 kN and 487 kN for impacts N1 and N2, respectively. The maximum deviations, sum-
 772 marized in Table 2 and ranging from -95% to $+450\%$, are estimated numerically by vary-
 773 ing each parameter by $\pm 50\%$. Figure 15a breaks down the contribution of each param-
 774 eter to the maximum uncertainty. Note that the relative errors are the same for both
 775 impacts N1 and N2. We can observe that a variation of impact speed v_n has the great-
 est effect on the impact force, followed by the rock density ρ .

Table 2. Hertz impact parameters^b

	v_c	α	v_n	δ_{max}	F_0	T_c	f_c
N1	15 m s^{-1}	15°	4 m s^{-1}	0.05 m	$(84 \begin{smallmatrix} +376 \\ -79 \end{smallmatrix}) \text{ kN}$	0.038 s	$(26 \begin{smallmatrix} +61 \\ -16 \end{smallmatrix}) \text{ Hz}$
N2	65 m s^{-1}	15°	17 m s^{-1}	0.16 m	$(487 \begin{smallmatrix} +2186 \\ -460 \end{smallmatrix}) \text{ kN}$	0.029 s	$(35 \begin{smallmatrix} +82 \\ -21 \end{smallmatrix}) \text{ Hz}$

^b Parameters for impacts N1 and N2: vertical impact speed v_c , angle α between the slope and the vertical, impact speed v_n normal to the slope, maximum indentation depth δ_{max} , impact force F_0 (with maximum deviations), contact time T_c , and corner frequency f_c (with maximum deviations).

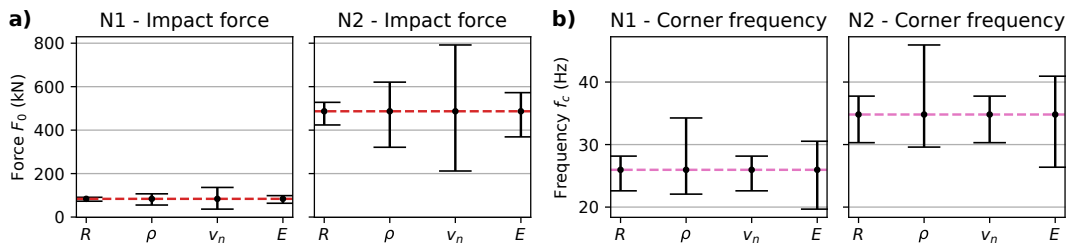


Figure 15. Influence of individual parameters on uncertainty of impact force F_0 (a) and corner frequency f_c (b). Impact parameters include boulder radius R , rock density ρ , normal impact speed v_n and Young's modulus E . Uncertainties are estimated numerically by varying each parameter by $\pm 50\%$. Note that the relative errors for N1 and N2 are identical.

776 The impact duration is estimated to be 0.038 s and 0.029 s for N1 and N2, respec-
 777 tively. As $f_c = 1/T_c$, it follows that the high-frequency content of the impacts are lim-
 778

779 ited by corner frequency 26 Hz and 35 Hz, respectively, with maximum deviations rang-
 780 ing from -62% to $+236\%$, see Table 2. The contribution from each parameter is indi-
 781 cated in Figure 15b for impacts N1 and N2. In contrast to the impact force, the frequency
 782 content is least sensitive to the normal impact speed v_n . It is most sensitive to rock den-
 783 sity ρ and Young’s modulus E .

784 An important result is that Hertz contact theory predicts a higher frequency con-
 785 tent for N2, related to the higher impact velocity. This agrees with the observed wave-
 786 forms in Figure 13d: impact N1 can hardly be detected in the high-frequency band (20-
 787 40 Hz), whereas impact N2 shows a clear pulse despite the slightly bigger source-receiver
 788 distance. The theoretical values agree well with the observations, predicting frequencies
 789 up to 26 Hz and 35 Hz for N1 and N2, respectively.

790 **5.3.2 Comparison of observed and synthetic waveforms**

791 Previously, we tried to associate pulses in the observed seismic signal to impacts
 792 observed in the video by crudely interpreting the signal after the time of impact. We will
 793 now use numerical simulations to obtain insights into travel times and expected wave-
 794 forms. As already mentioned, the intention is not to reproduce observed waveforms, but
 795 rather to understand which signal characteristics can be ascribed to a single impact.

796 Observed and synthetic signals are compared in the frequency band of 10-20 Hz,
 797 in which we identify short signal pulses caused by the rockfall impacts. At the same time,
 798 20 Hz constitutes the upper frequency limit of our simulations. In order to ensure com-
 799 parability, the observed signals are corrected with the site amplification functions cal-
 800 culated in Appendix C and subsequently convolved with the 7 Hz Ricker wavelet used
 801 in the simulations. The simulated signals are calibrated with the maximum impact forces
 802 for N1 and N2 estimated above using Hertz theory.

803 Initially, observed and synthetic seismograms are normalized for easier compari-
 804 son of the waveforms. Figure 16 compares vertical ground velocity recorded at stations
 805 BON, BOR, DSO, and SNE with simulations from impacts N1 and N2 for models with
 806 a flat surface and with Dolomieu topography. Source positions of the two impacts are
 807 estimated from the videos, see Figure 13b. As the exact source direction of the real im-
 808 pacts are unknown, the variability of synthetic waveforms is shown for different force di-
 809 rections, namely a vertical force F_z , a force F_n normal to the slope and a force F_t tan-
 810 gential to the slope.

811 Analyzing the synthetic seismograms, it can generally be observed that N1 produces
 812 smaller amplitudes compared to N2. This is due to the estimated impact forces of 84 kN
 813 and 485 kN, respectively (see Table 2).

814 While seismograms from the model with a flat surface keep approximately the same
 815 relative amplitudes between N1 and N2 at all different stations, seismograms from the
 816 model with topography show more variability. For example, at station BON, the am-
 817 plitudes of impact N1 are much bigger for the model with topography than for the flat
 818 model. This corresponds better to the real observations, where the maximum amplitude
 819 of impact N1 is comparable to the maximum amplitude of impact N2. In contrast, at
 820 station DSO, the signal of impact N1 is very small as opposed to the signal of N2 for the
 821 model with topography. Again, this corresponds well with the observations. As impacts
 822 N1 and N2 are located very close to each other, the relative amplitudes on the flat model
 823 are mainly determined by the relative impact force. In contrast, surface topography in-
 824 fluences both the relative (vertical) source position and the propagation path, so that
 825 small source displacements can cause local amplification or deamplification of the sig-
 826 nal. Measuring the signal at station DSO, the source moves from a deamplified zone at
 827 N1 towards an amplified zone at N2, as suggested by the reciprocal amplification pat-
 828 tern provided in the Supporting Information.

829 From the simulations with a flat surface, three wave packets can be observed fol-
 830 lowing each impact, which are well separated from each other on the more distant sta-
 831 tions BOR, DSO, and SNE. These three wave packets correspond to a body wave, a first-

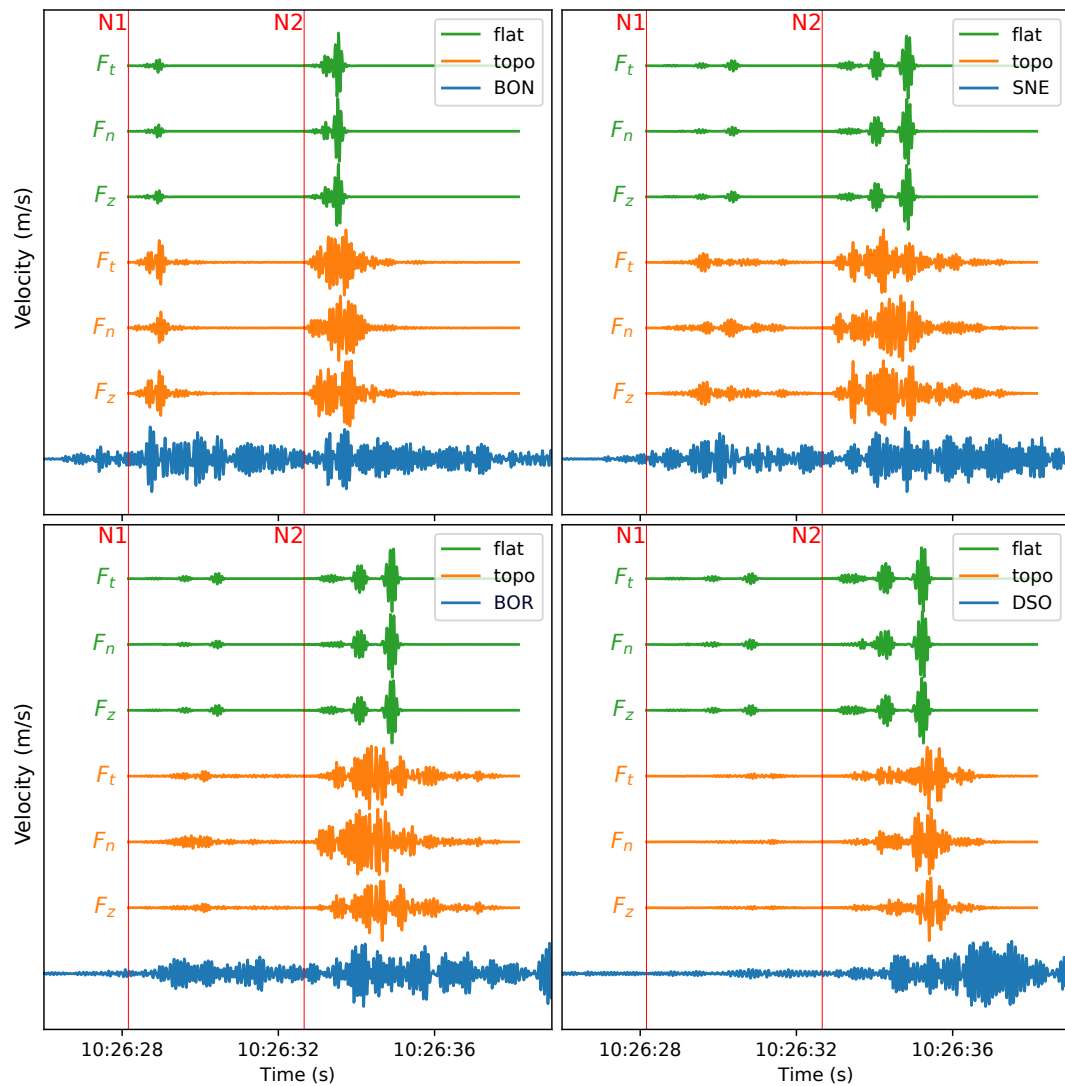


Figure 16. Comparison of recorded rockfall signal (blue, for a time window as in Figure 13d) with synthetic waveforms of impacts N1 and N2 for the model with a flat surface (green) and with Dolomieu topography (orange). The variability of the synthetic waveforms is demonstrated for a vertical force F_z , a force F_n normal to the slope and a force F_t tangential to the slope. Red vertical lines indicate impact times N1 and N2 from the video. Corresponding source positions are shown in Figure 13b. All signals show vertical ground velocity and are normalized by their maximum. SEM configurations correspond to 15 and 16 in Table A1.

832 mode Rayleigh wave, and a fundamental mode Rayleigh wave (see e.g. Figure 3). The
 833 arrival time of the first-mode Rayleigh wave is in good agreement with the first major
 834 pulse after each impact. This suggests that the Lesage generic velocity model represents
 835 the shallow subsurface velocity around Dolomieu crater reasonably well. However, for
 836 the flat model, the amplitude of the first-mode Rayleigh wave is consistently smaller than
 837 the amplitude of the fundamental mode. A corresponding amplitude variation cannot
 838 be identified on the real signals. In contrast, simulations on the model with topography
 839 generate more complex waveforms. This increased complexity corresponds better to the
 840 observed signals, even if the waveforms do not fit perfectly. The variation of the force
 841 direction modifies the waveforms more than in the flat case. Also, waveforms vary greatly

842 from station to station. This is not observed in the flat case, in which the waveforms are
 843 very similar for stations at comparable source-receiver distances (i.e. BOR, DSO, and
 844 SNE). Similar observations are made for synthetic signals generated by neighboring source
 845 positions (10 m spacing), exhibiting much more variability for the case of topography,
 846 see Supporting Information.

847 It cannot be excluded that the observed amplitude variations and waveform com-
 848 plexity can be explained by heterogeneous velocity structures or the superposition of mul-
 849 tiple (here undetected) sources. Nonetheless, the simulations indisputably show that these
 850 effects are caused by surface topography, for which high-resolution models are available.
 851 Consequently, surface topography must be taken into account when interpreting high-
 852 frequency rockfall seismic signals and multiple signal pulses must not directly be inter-
 853 preted to be caused by multiple impacts.

854 Finally, observed and synthetic seismograms are compared without normalization.
 855 In this way, the absolute signal amplitudes calibrated by the Hertz impact force are eval-
 856 uated. The total value of the acting force as well as its direction are determined by a vec-
 857 tor sum of the forces normal and tangential to the slope. Tangential force F_t is inferred
 858 from the maximum normal impact force $F_n = F_0$ assuming Coulomb friction $F_t = \mu F_0$,
 859 where μ is the material-specific friction coefficient. We assign $\mu = 0.7$, a typical value
 860 used for rockfalls at Dolomieu crater (e.g. Hibert, Mangeney, et al., 2014). The result-
 861 ing signal amplitudes for model simulations with a flat surface and with topography are
 compared with the observed rockfall signals in Figure 17.

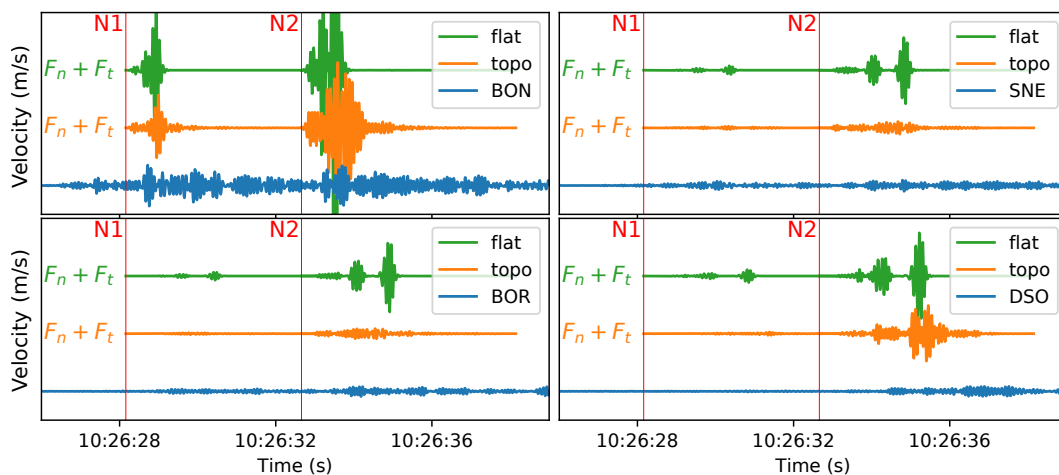


Figure 17. Amplitude comparison of recorded rockfall signal (blue, for a time window as in Figure 13d) with synthetic waveforms of impacts N1 and N2 for the model with a flat surface (green) and with Dolomieu topography (orange). The synthetic seismic source is constructed by summing force F_n and F_t normal and tangential to the slope. F_n corresponds to the maximum impact force F_0 as shown in Table 2. Red vertical lines indicate impact times N1 and N2 from the video. Corresponding source positions are shown in Figure 13b. All signals show vertical ground velocity and are normalized by their maximum. SEM configurations correspond to 15 and 16 in Table A1.

862 Despite the uncertainties on impact force F_0 and the assumptions on elastic contact
 863 and Coulomb friction, which are certainly violated by a certain amount of plasticity
 864 and surface roughness, we are able to achieve order-of-magnitude agreements with
 865 the observed signals. The application of the Hertz contact theory to calibrate the sim-
 866 ulations therefore offers a solution to this poorly constrained problem.
 867

868 When comparing the amplitudes between the model with a flat surface and the model
 869 with topography, one notices that the relative amplitude changes between the stations.
 870 While amplitude decreases linearly with the distance from the source for the flat model,
 871 this is not the case for simulations with topography. For example, for the flat model, the
 872 maximum amplitude decreases drastically from station BON to station DSO. In contrast,
 873 for the model with topography, the signal amplitude of impact N2 at station DSO is only
 874 about half that of station BON, which can be assigned to topographic amplification.

875 **6 Conclusion**

876 Using Spectral-Element-based simulations with high-resolution surface topography,
 877 we were able to broadly explore the influence of topography on the propagation of wave
 878 fields generated by surface point loads and to achieve order-of-magnitude agreement with
 879 seismic measurements from rockfalls at Dolomieu crater.

880 In the first half of the present study we numerically simulated the topography ef-
 881 fect for different earth models. It was shown that the topography effect is significantly
 882 altered depending on the underlying velocity model. For example, for a homogeneous
 883 model, seismic energy is easily directed downwards by the topography into the subsur-
 884 face. This is contrary to a velocity-depth profile with a strong gradient, as proposed by
 885 Lesage et al. (2018) for the shallow velocity structure of volcanoes, for which more seis-
 886 mic energy is trapped close to the surface leading to more complex spatial distributions
 887 of amplification.

888 Studying the geometric features of Dolomieu-like crater topographies suggests that
 889 curvature variations affect the seismic wave field more than depth variations. However,
 890 the locally inherent complexity of the topography prohibits a holistic generalization of
 891 its effect on the seismic wave field. Nevertheless, our approach provides a methodology
 892 to quantify the influences of the topography. This methodology can be used to other study
 893 sites by changing the model parameters regarding the domain surface and the medium
 894 properties, which are inevitably coupled. In this respect, scattering from 3D soil hetero-
 895 geneities, which can strongly disturb wave propagation, must be considered. Regarding
 896 the study site at Dolomieu crater, previous studies and our findings indicate that scat-
 897 tering does not dominate the signals. Local site amplification at the stations could be
 898 taken into account by using amplification factors estimated from VT events, which sig-
 899 nificantly improved the agreement between simulations and observations. In general, how-
 900 ever, it cannot be guaranteed that surface waves will receive the same amplification as
 901 a vertically incident wave field and it is recommended that wave propagation, including
 902 subsurface heterogeneities, be modeled when the necessary information is available.

903 In the second half of the present study, the simulations were compared with mea-
 904 surements from rockfalls at Dolomieu crater. Different rockfalls with similar impact lo-
 905 cations were investigated by calculating spectral ratios between the stations, thereby ex-
 906 tracting the signature of the seismic source. The agreement between the observed spec-
 907 tral ratios indicates identical path effects that were reproduced by simulations using the
 908 model with topography. It was further shown that the spectral ratios are dominated by
 909 propagation along the topography rather than by the direction of the seismic source. As
 910 the latter is hard to estimate precisely, this finding can have practical applications, for
 911 example for the locating of rockfalls.

912 Single impacts were shown to be able to generate complex waveforms with mul-
 913 tiple pulses depending on medium properties and topography. For topography, variations
 914 of source direction and position can strongly modulate the waveforms.

915 Using Hertz contact theory, signal features were linked to impact parameters. Es-
 916 timating the maximum impact force helped to calibrate the simulation and achieve order-
 917 of-magnitude agreement with observed signal amplitudes. Associating higher frequen-
 918 cies with increased impact speed explained the observed frequency content of rockfall
 919 impacts.

920 The combination of Hertz contact theory and wave propagation simulations is an
 921 important step for the interpretation of rockfall seismic signals based on the underlying
 922 physical processes. The Hertz impact theory is frequently used to predict the impact force
 923 of rockfalls, for example for the design of protective structures (e.g. Volkwein et al., 2011).
 924 Also, laboratory experiments show the validity of Hertz theory concerning the waves gen-
 925 erated by the collision of a ball on a massive plate (e.g. McLaskey & Glaser, 2010) or
 926 grains on a plate (Bachelet et al., 2018). However, only a few studies apply the theory
 927 to seismic signals from real-scale rockfalls (Farin et al., 2015; Hibert, Malet, et al., 2017).
 928 A limiting factor is the complexity of the rockfall source that often consists of multiple
 929 simultaneous impacts. For this reason, application to artificially triggered rockfalls, en-
 930 suring separate impacts of a single boulder, would help validate the Hertz theory in the
 931 field and enhance our understanding of real impact processes.

932 **Appendix A Configuration of SEM simulations**

933 Table A1 contains all configurations of the SEM simulations conducted for the present
 paper and indicates the section in which each is used.

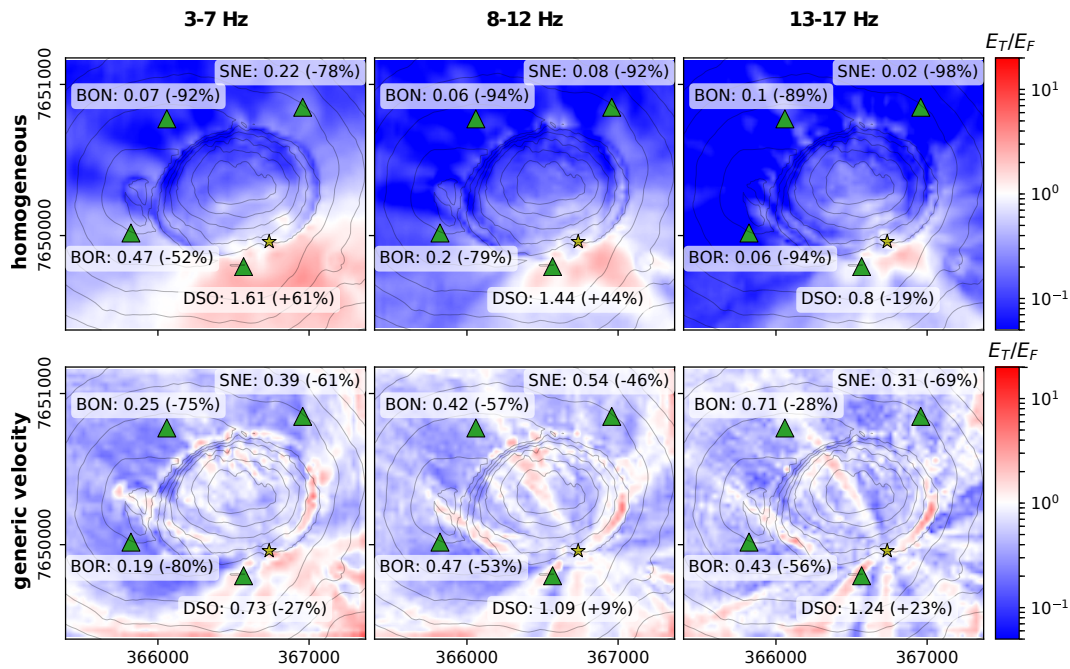
Table A1. Configuration of SEM simulations and usage in present article^c

	Velocity model	Surface	α	Source	Output	Sections
1	homogeneous	Flat	Yes	R1; <i>Z</i>	Seismograms	4.1
2	homogeneous	Topo 20m	Yes	R1; <i>Z</i>	Seismograms	4.1
3	homogeneous	Topo 20m	No	R1; <i>Z</i>	Snapshots	3.4
4	low- v_S layer	Flat	Yes	R1; <i>Z</i>	Seismograms	4.1
5	low- v_S layer	Topo 20m	Yes	R1; <i>Z</i>	Seismograms	4.1
6	low- v_S layer	Topo 20m	No	R1; <i>Z</i>	Snapshots	3.4
7	generic	Flat	Yes	R1; <i>Z</i>	Seismograms	3.3, 4.1
8	generic	Topo 20m	Yes	R1; <i>Z</i>	Seismograms	3.3, 4.1
9	generic	Flat	Yes	R1; <i>N</i>	Seismograms	4.2
10	generic	Topo 20m	Yes	R1; <i>N</i>	Seismograms	4.2
11	generic	Topo 20m	No	R1; <i>Z</i>	Snapshots	3.4
12	generic	Flat	Yes	S1; <i>Z</i>	Seismograms	4.3
13	generic	Flat rough	Yes	S1; <i>Z</i>	Seismograms	4.3
14	generic	Craters	Yes	S1; <i>Z</i>	Seismograms	4.3
15	generic	Flat	Yes	reciprocal	Seismograms	5.2.1, 5.3.2
16	generic	Topo 10m	Yes	reciprocal	Seismograms	3.3, 5.2.2, 5.3.2

^c Configurations according to velocity model, surface characteristics, attenuation ‘ α ’, source position and direction, and output, along with indication of the section of the present article in which each configuration is referred to. ‘Topo 20m’ and ‘Topo 10m’ refers to topography with 20 m and 10 m resolution, respectively. ‘R1’ corresponds to source position of rockfall 1, Figure 1b; ‘S1’ to the source position of the synthetic crater study, Figure 7. Reciprocal simulations are carried out for each seismometer (BON, BOR, DSO, SNE) and for each direction (*E*, *N*, *Z*).

935 **Appendix B Energy amplification in different frequency bands**

936 Figure B1 shows energy amplification in three different frequency bands for the ho-
 937 mogeneous velocity model (top) and the generic velocity model (bottom). Rayleigh wave-
 938 lengths in frequency band 3-7 Hz of the homogeneous model ($\lambda \approx 1000 \text{ m s}^{-1} \div 10 \text{ Hz} =$
 939 100 m) are comparable to those in frequency band 8-12 Hz of the Lesage generic model
 940 ($\lambda \approx 580 \text{ m s}^{-1} \div 5 \text{ Hz} \approx 116 \text{ m}$, see dispersion curves in Figure 2c). However, we can
 941 observe that the amplification patterns differ in these two frequency bands. This sug-
 942 gests that the respective amplification patterns are not only characteristic of a certain
 943 wavelength. The wave propagation essentially depends on the velocity model which hence
 results in different amplification patterns.



944 **Figure B1.** Amplification of total kinetic energy in frequency bands 3-7 Hz (*left*), 8-12 Hz
 945 (*middle*) and 13-17 Hz (*right*) for the homogeneous velocity model (*top*) and the Lesage generic
 946 velocity model (*bottom*). The yellow star denotes the source and the green triangles the sta-
 tions. Annotations indicate ratios measured at the station locations as well as the percentage of
 topographic amplification. Neighboring contour lines differ by 60 m in elevation.

944

945 **Appendix C Estimation of site effects caused by local subsurface struc-**
 946 **tures**

947 Before calculating the site amplification functions, Figure C1a compares spectral
 948 ratios from the uncorrected rockfall signals to those from simulations with the models
 949 using the Dolomieu topography. Note that the ratios from the simulations seem to be
 950 smaller than the real values. In particular ratio SNE/BON is strongly underestimated,
 951 especially for the horizontal components. This is possibly caused by local structures in
 952 the subsurface which are not accounted for in the simulations. Local site effects are es-
 953 timated to correct the recorded signals and ensure comparability between observations
 954 and simulations.

955 Site effects are estimated from seismic signals generated by volcano-tectonic (VT)
 956 events which are centered around 2 km below Dolomieu crater. Thirty-six events are se-

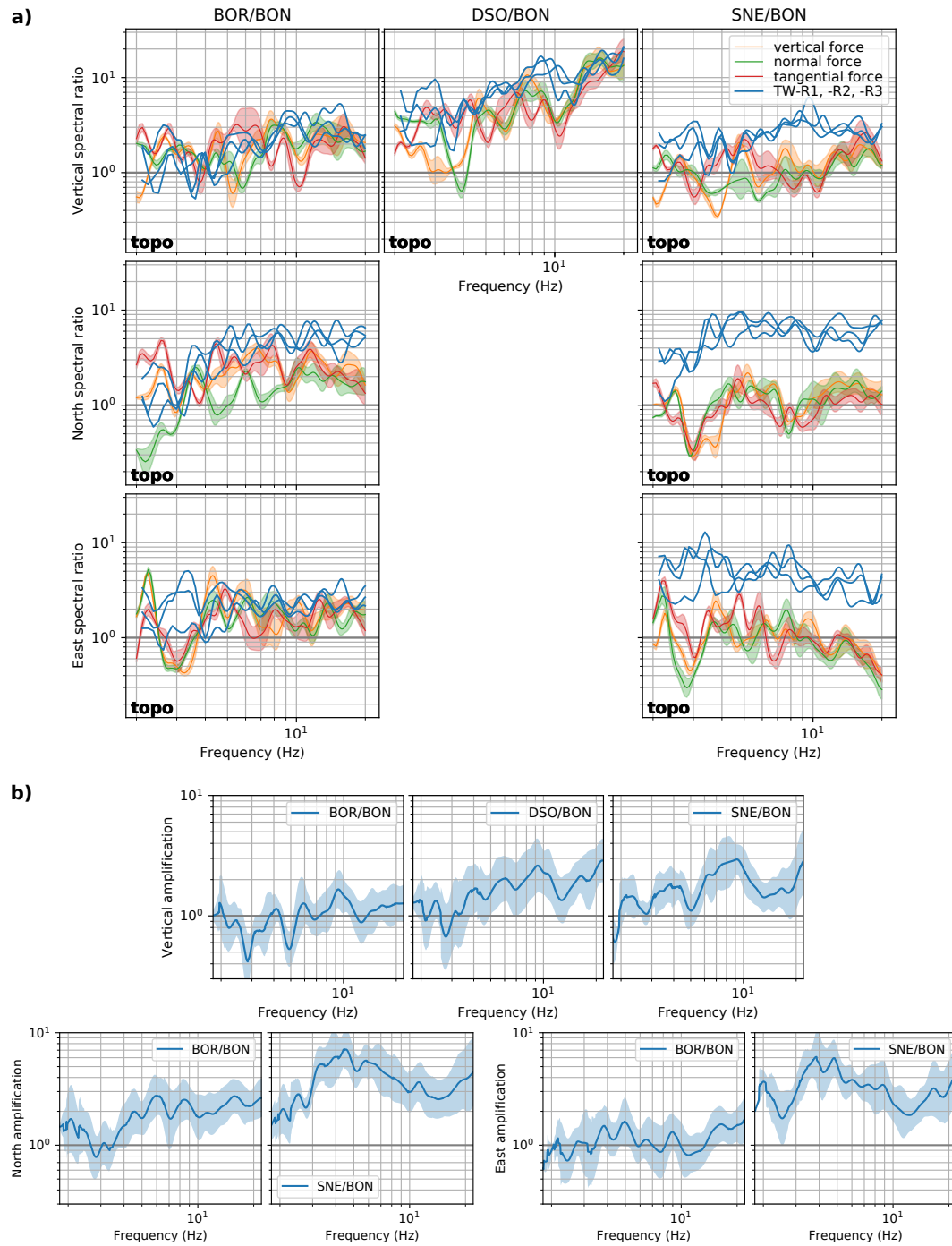


Figure C1. a) Comparison of spectral ratios as in Figure 12 but with before site effect correction of the recorded signals (blue). This leads to partly strong deviations between observations and simulations. b) Spectral amplification functions relative to reference BON for ground velocity of vertical (*top*), north (*bottom left*) and east (*bottom right*) component. The blue-shaded zone indicates the standard deviation of the amplification distribution from all VTs.

957
958

lected from a catalog compiled by Duputel et al. (2019). To compute the amplification functions, BON was qualified as an adequate reference station based on the low spec-

959 tral amplitudes of both VT signals and H/V noise ratios. The spectral ratios are com-
960 puted from FFT spectra after applying the smoothing function proposed by Konno and
961 Ohmachi (1998) to avoid spurious fluctuations. Figure C1b shows the mean spectral am-
962 plification functions and their standard deviation calculated from all VT events for all
963 components.

964 Strongest amplification is experienced by station SNE with factors up to 7 on its
965 horizontal components. This can explain the strong mismatch between observations and
966 simulations that are visible in Figure C1a. The vertical component of single-component
967 station DSO also seems to be amplified with a peak around 9 Hz. Less evidence of am-
968 plification is found for station BOR, except for its north-component which is amplified
969 by a factor of 2 for frequencies above 5 Hz.

970 **Appendix D Observed and simulated spectral ratios for rockfall sources** 971 **in the southwest**

972 To reinforce the findings of section 5.2.2 that spectral ratios are characteristic of
973 the source position and can be reproduced when surface topography is taken into account,
974 the same analysis is carried out here for rockfall sources located on the southwestern crater
975 wall. Snapshots taken from camera CBOC of the three observed rockfalls are shown in
976 Figure D1, together with the generated seismic signals recorded on the vertical compo-
977 nents. For the times of the shown images, marked on the seismic signals by the verti-
978 cal dotted red lines R1, R2, and R3, all the rockfalls are located in the same area.

979 The camera images reveal that each of the rockfalls involves at least two boulders
980 moving downslope simultaneously. While the boulders of rockfall **1** originate from be-
981 low the camera position, boulders of rockfall **2** and **3** come from the right-hand border
982 of the image. At the time of the snapshot, the trajectories of the three rockfalls cross.
983 From a window of ± 4 s around this time, the spectral ratios are computed from the ob-
984 served signals and shown in Figure D2 (blue, site-effect corrected). As for the rockfalls
985 analyzed above, the spectral ratios from the three events in the southwest are very sim-
986 ilar to each other across the whole frequency range for all station pairs.

987 The spectral ratios are now compared to simulations using the model with Dolomieu
988 topography. As above, three input force configurations are tested (i.e. a vertical force,
989 a force normal to the slope and a force tangential to the slope in direction of the strongest
990 gradient) to investigate the dependency of the ratios on the source direction.

991 We can generally observe that the simulated spectral ratios agree very well with
992 the observed spectral ratios. Changing the source direction does not essentially influence
993 the spectral ratios, except for frequencies below 5 Hz, which is similar to the observations
994 in Figure 12. The similarity at higher frequencies suggests that the ratios are dominated
995 by propagation along the topography rather than by the source mechanism. The strongest
996 deviation between observations and simulations at high frequencies is visible on ratio SNE/BON
997 for the east-component. In comparison with the observed spectral ratios, the simulated
998 amplitudes measured at station SNE are strongly underestimated with respect to sta-
999 tion BON. This could be caused either by soil heterogeneities, which are not considered
1000 in the simulations, or by uncertainties of the source position and the fact that the rock-
1001 fall contains at least two boulders which simultaneously impact the ground.

1002 Analysis of the rockfalls located in the southwestern part of Dolomieu crater sup-
1003 ports the findings of section 5.2.2 indicating that the spectral ratios are characteristic
1004 of the source location, and can be reproduced by taking into account the surface topog-
1005 raphy, while source direction is not dominant, in particular at high frequencies.

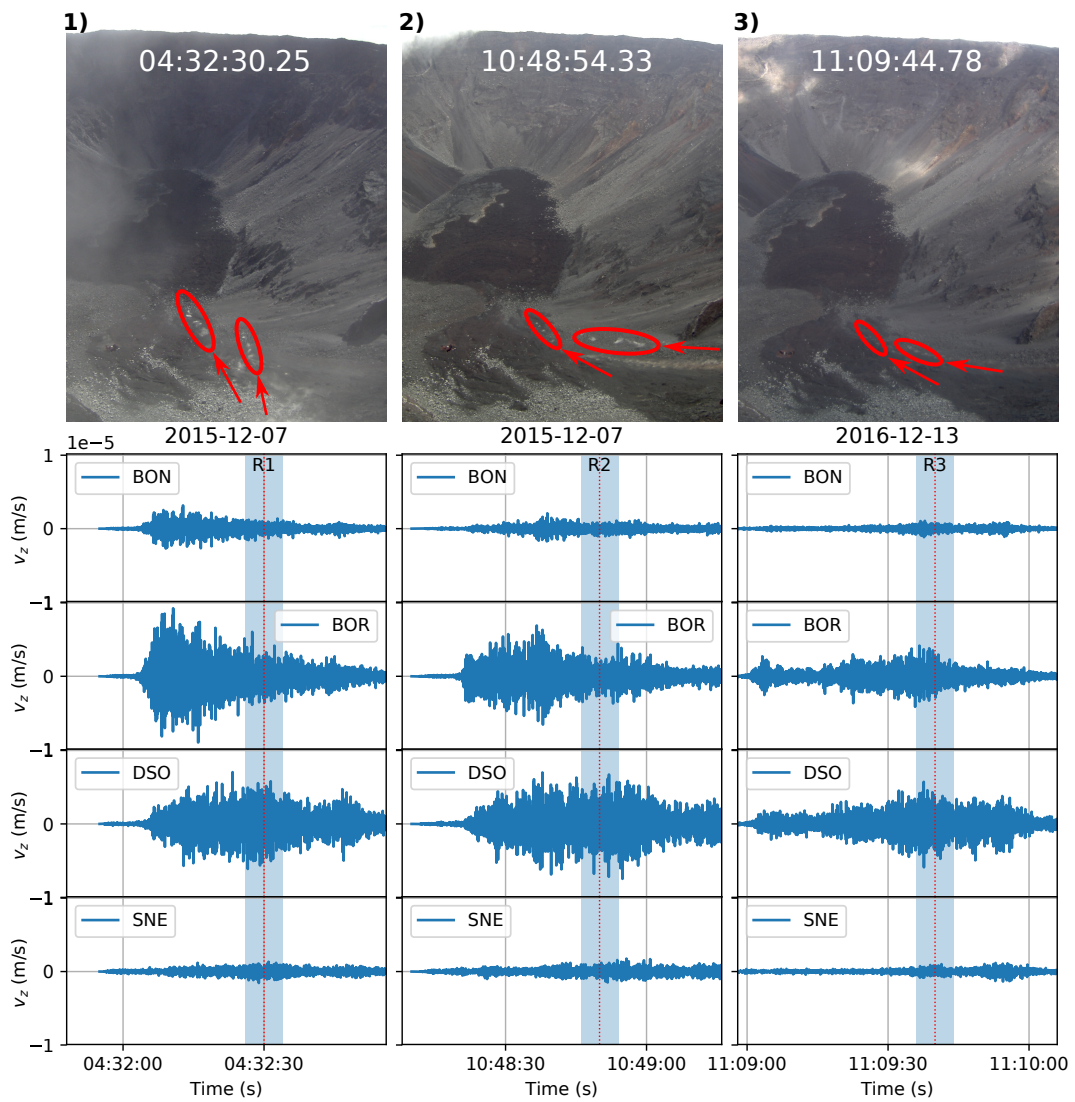


Figure D1. *Top:* Snapshots taken from camera CBOC of three rockfalls times for which all the rockfalls are in comparable locations. Positions and directions of the boulders are indicated by red circles and arrows. The trajectory of the rockfall on 13 December, 2016 is indicated as event **2** on the map in Figure 1b *Bottom:* Corresponding seismic signals (vertical velocity). The vertical dotted lines R1, R2 and R3 mark the time of the camera snapshot shown above. The blue-shaded zones display the time windows of ± 4 s around R1, R2, and R3 in which spectral station ratios of the signals are computed.

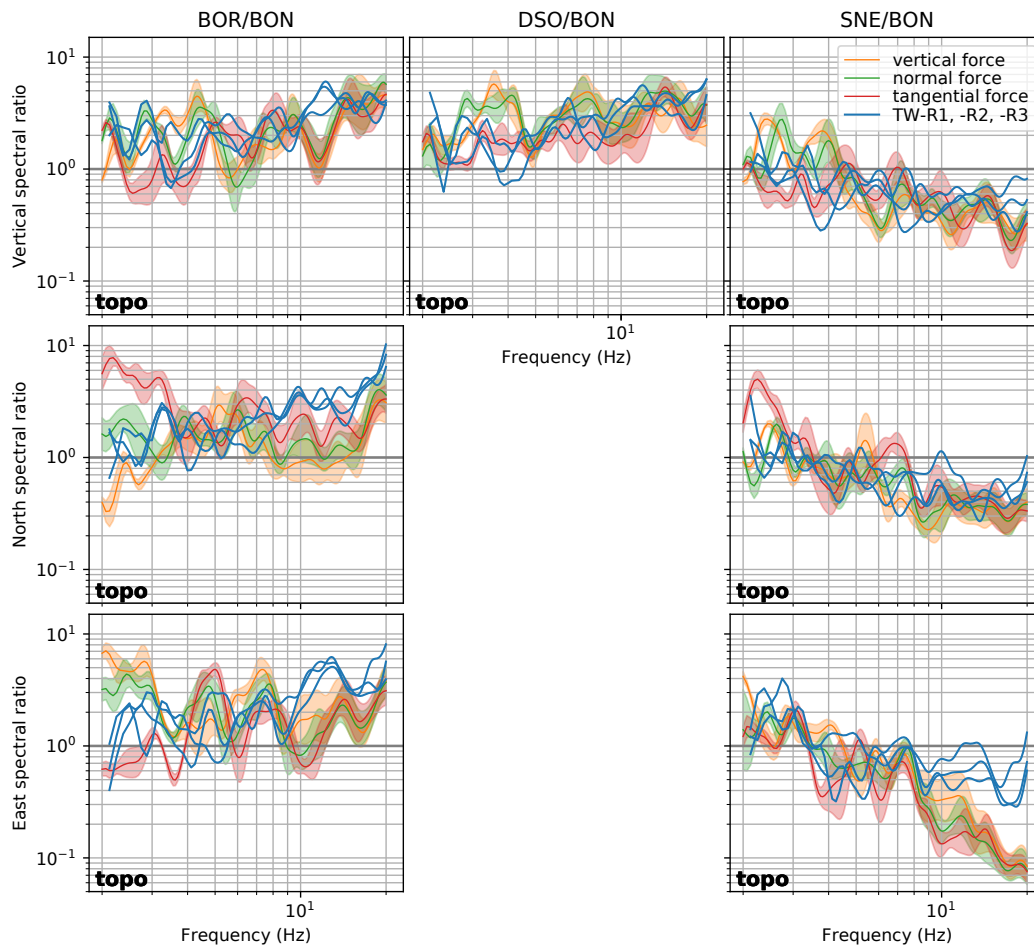


Figure D2. Spectral ratios BOR/BON, DSO/BON and SNE/BON calculated from real signals TW-R1, -R2, -R3 (as defined in Fig. D1) and from simulations on the domain with Dolomieu topography for the vertical- (*top*), the north- (*middle*), and the east- (*bottom*) component. Simulations are carried out with varying source directions: vertical force, force normal to the slope and force tangential to the slope. The shaded zones of the simulated ratios indicate the standard deviation around the mean value of seventeen neighbouring source positions located close to index number 2 on the trajectory of event 2 on the map in Figure 1b.

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 1013 Paris (IPGP) via the VOLOBISIS Portal: <http://volobsis.ipgp.fr/query.php>. Camera data
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