Numerical and observational study of *Sn*-to-*Lg* conversion due to crustal-thickening: implications for identification of continental mantle earthquakes

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Key points

- Synthetics and data show the *Sn/Lg* method successfully identifies mantle earthquakes with thickening crust across the Himalaya.
- *Sn*-to-*Lg* conversions can be recognized by enhanced high frequency content of *Lg*
- *Lg* frequency content discriminates between crustal and mantle near-Moho earthquakes

Abstract

We study *Sn*-to-*Lg* conversion at regional distances due to significant crustal thickening, particularly in the context of using *Sn* and *Lg* amplitude ratios (*Sn/Lg*) to identify continental mantle earthquakes. We further enhance recent developments in computational seismology to perform 2.5D simulations up to 5 Hz and 2,000 km. Our simulations compare propagation in a reference, constant-thickness crust from a source at three depths straddling the Moho, to 48 models of the same three sources propagating through Moho ramps of four different widths (dips) at four different distances from the source. We compare our synthetics to data from 12 earthquakes recorded on the HiCLIMB array across Tibet, of which six events from northwestern Tibet traverse no major crustal-thickness variation, and six located south of the Himalaya cross a major Moho ramp. Our observations on real data show that amplitude perturbations on individual *Sn* and *Lg* waves are smooth and mostly limited to near the ramp end. Even the more-pronounced amplitude variations seen in our simulations show that *Sn/Lg* for mid-crustal earthquakes is consistently lower than those for mantle earthquakes. Hence we can directly compare *Sn/Lg* for ramp-crossing and non-ramp-crossing earthquakes and identify new mantle earthquakes in northern India. *Sn*-to-*Lg* converted waves may be readily detected near the Moho ramp end through an enhancement in high-frequency content. In addition, we observe higher frequency content in *Lg* from crustal than from mantle earthquakes, which offers a new discriminant for continental mantle earthquakes based on frequency content of *Lg* waves alone.

89   **Plain language summary**

90   Seismic waves *Sn* and *Lg* respectively propagate largely below and above the Moho. Previous

91   work showing that *Sn* and *Lg* amplitudes can distinguish whether near-Moho continental

92   earthquakes nucleated in the crust or mantle (the '*Sn*/*Lg* method') used only 1D (flat-Moho)

93   theory and synthetics, and data from areas with little Moho topography. Here we extend this

94   work with synthetic seismograms across large Moho ramps and with data recorded across the

95   Himalaya from India to Tibet. By comparing earthquakes with source-receiver raypaths that do

96   and do not cross a Moho ramp we show the *Sn*/*Lg* method can still identify mantle earthquakes

97   provided multiple recorders are used. We also show that the frequency content of *Lg* contains

98   information about *Sn*-to-*Lg* conversions, and can by itself be used to identify mantle earthquakes.

99   Traditionally, *Sn* and *Lg* waves have not been modeled at high-frequencies (>1 Hz) and long-

100   distances (>1000 km) due to high computing costs. Here, we take advantage of and enhance

101   recent developments in computational seismology to model *Sn* and *Lg* propagation up to 5 Hz

102   and for 2000 km through a 2D lithosphere, paying special attention to their amplitude ratio and

103   its application to distinguish exotic continental mantle earthquakes from commonplace crustal

104   earthquakes.

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1. Introduction

Seismic waves *Sn* and *Lg* are the most prominent arrivals on high-frequency (~1–5 Hz) seismograms recorded at regional distances (~200–2,000 km). They are guided shear waves within the entire crust (*Lg*) or the entire lithosphere (*Sn*), and can be represented equivalently either by Airy phases from surface-wave normal modes (Stephens and Isacks,1977, Knopoff, 1973) or by interference patterns of waves multiply reflected between the surface and the Moho top-side (Oliver and Ewing, 1958) or under-side (Červený and Ravindra, 1971; Menke and Richards, 1980), respectively (Fig. 1). Their excitation and propagation characteristics derived from the above representations are directly related to the wave amplitudes that have been useful for a variety of purposes such as determining focal depths for crustal earthquakes from amplitude spectra (Baker et al., 2004), serving as the dominant measure for regional earthquake magnitude (e.g. Patton and Walter, 1993), monitoring nuclear tests based on *Pg* and *Lg* amplitude ratios (e.g. Zhang and Wen, 2013), as well as estimating local properties relating to the attenuation (e.g. Mousavi et al., 2014) and amplification (i.e. seismic hazards, e.g. Kebeasy and Husebye, 2003, Rodgers et al., 2019, 2020) of these waves.

Recently, addressing a half-century-long controversy regarding whether earthquakes can nucleate in the continental mantle (Chen and Molnar, 1983; Maggi et al., 2000; Chen and Yang, 2004; Schulte-Pelkum et al., 2019; Priestley et al., 2008; Craig et al., 2011; Prieto et al., 2017), we demonstrated the use of *Sn* and *Lg* amplitude ratios (hereafter "*Sn/Lg*") to discriminate continental mantle earthquakes from crustal ones using Tibetan earthquakes recorded on the Tibetan plateau (Wang and Klemperer, 2021) (Fig. 2). The signature of a mantle origin is a higher *Sn/Lg* compared with nearby crustal earthquakes recorded on a common array. For a group of earthquakes in NW Tibet, *Sn/Lg* ratios > 2 (averaged over many stations) were found to identify sub-Moho earthquakes. This method has the advantages of making the discrimination by relying on prominent waveform features of the earthquakes themselves (as opposed to Zhu and Helmberger, 1996 and Yang and Chen, 2010, who relied on more subtle waveform features), thus avoiding comparing independently derived earthquake and Moho depths at different locations, which has been a popular method (Chen and Yang, 2004, Priestley et al., 2008), and also can be performed using any stations/arrays that lie within regional distances of an

149   earthquake (as opposed to Schulte-Pelkum et al., 2019 who relied on stations essentially on top

150   of earthquakes).

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152   Our *Sn/Lg* method is based on predictions from 1D surface-wave normal-mode theory, but given

153   the *Sn* and *Lg* interconversions (*Lg* blockage or leakage) created when waves are incident on a

154   dipping Moho, it is far from certain how the method will perform if there exists a large-scale

155   structural variation between the earthquakes and recording stations (e.g. earthquakes in India

156   recorded by stations in Tibet). Necessary corrections may be small – Song and Klemperer (2023)

157   show general agreement between the catalog depths and *Sn/Lg* of hundreds of earthquakes with

158   paths crossing the boundaries of Tibetan Plateau recorded on either of two permanent stations

159   (KBL and LSA) – or may be significant, as where *Lg* blockage is used to study large-scale

160   geologic features (e.g. North Sea: Mendi et al.,1997, Japan: Furumura et al., 2014, Pyrenees:

161   Sens-Schönfelder et al., 2009).

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163   Given the number of seismological applications utilizing regional-wave amplitudes, and that

164   large-scale Moho topography is often well-known, many attempts have been made to quantify *Lg*

165   blockage as *Lg* propagates through a suddenly thinned crust. Coupled-mode theory builds on the

166   1D surface-wave eigenproblem which can synthesize regional waves with only vertical (1D)

167   heterogeneity. Coupled-mode theory represents the wavefield as a sum of basis functions

168   (motion-stress vectors for a 1D problem, e.g. Aki and Richards, 2002) with laterally-varying

169   amplitude coefficients obtained through the orthogonality principle of the normal modes

170   (Maupin, 1988). Most relevant here are to consider the width across which Moho depth varies

171   (Kennett, 1972; Drake, 1972; Kennett 1984; Maupin, 1988) and to incorporate undulating

172   structural boundaries using local modes (i.e. motion-stress vectors corresponding to a flat (1D)

173   model locally identical to a small section of the laterally varying 2D model; Odom, 1986) and

174   representing the continuity conditions on the tilted surfaces as a volume force in both 2D

175   (Maupin, 1988) and 3D (Tromp, 1994). A 2D coupled-local-mode method, incorporating all

176   these ideas, was applied to *Lg* propagation in the North Sea (Maupin, 1989) to model transmitted

177   and reflected wavefields for incident waves both perpendicular and at a sub-critical angle to the

178   strike of the Moho topography. Maupin (1989) reported little difference between perpendicular

179   and oblique incidences; the reflected wavefield is negligible and the strong *Lg* attenuation seen in

180   the North Sea cannot be fully explained simply by structural effects, a conclusion that has been

181   corroborated by later studies (Cao and Muirhead, 1993; Mendi et al., 1997) using 2D finite-

182   difference simulations. An important observation is that mode-coupling occurs most strongly

183   between neighboring modes. In Maupin (1989)'s North Sea model at a fixed frequency of 1 Hz,

184   *Lg* mostly leaks into the mantle as *Sn* waves from the first (lowest) few *Sn*-forming normal

185   modes (Fig. 3), as predicted by Kennett (1984). This means that only the lowest few *Sn*-forming

186   normal modes, or the highest few *Lg*-forming normal modes, get enhanced by *Sn* and *Lg*

187   interconversion, and if these enhanced modes do meaningfully contribute to either the *Sn* or *Lg*

188   wavetrain then they contribute more to the low-frequency content of *Sn* or the high-frequency

189   content of the *Lg* wavetrain. An alternative to the coupled-mode method is the ray-diagram

190   method (Kennett, 1986 for *Lg*; Xie, 1996 for *Pn*), whose results are mostly graphical and do not

191   account for interference between different rays once their initial coherent pattern is broken

192   (Kennett, 1986). Nonetheless, for an initial bundle of rays with the same inclination (i.e. apparent

193   velocity), focusing and de-focusing effects due to the lateral structure can be clearly seen (e.g.

194   Kennett, 1986, his figures 2 and 3). These methods study the interactions of different modes (i.e.

195   different dispersion relations: frequency vs. wavenumber) by either fixing the frequency (the

196   coupled-mode methods) or the wavenumber (proxy to apparent velocity, the ray-diagram

197   method). These methods yield valuable insights, but cannot represent the full broadband

198   wavefield, which for regional waves is dominated by interference patterns.

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200   Fully-numerical simulations can calculate the full broadband wavefield for any arbitrary

201   structure, however, significant computational challenges exist given the frequency and range

202   requirements for simulating regional waves. We are not aware of any 3D simulations that

203   simultaneously reach frequencies up to 5 Hz and distance ranges up to 2,000 km, common

204   observational parameters for *Sn* and *Lg* waves. Furumura et al. (2014) simulated regional wave

205   propagation around Japan up to 1.5 Hz; and Rodgers et al. (2019, 2020) simulated ground motion

206   in the San Francisco Bay Area covering an area of 120 km x 80 km up to 10 Hz. More

207   importantly, these 3D simulations are run with very specific models, so are hard to generalize to

208   other cases. On the other hand, 2D simulations, which recently focused on *Pn* propagation (Bakir

209   and Nowack, 2012; Xie and Lay, 2017a&b; Wang et al., 2017), are attractive as they are much

210   cheaper, so may simultaneously satisfy the frequency and range requirements, and may be more

211   generalizable. However, these simulations, if performed in a Cartesian grid, require the earth-

212   flattening transformation to produce physical sphericity which is vital for simulating interference

213   head waves such as *Pn* and *Sn*. More importantly, these 2D simulations require a non-

214   straightforward correction from their 2D line sources to 3D point sources (Li et al., 2014), and

215   this correction cannot be exact if lateral heterogeneities exist (Li et al., 2014).

216

217   Here we establish first-order features of *Sn* and *Lg* transmission and inter-conversion with a set

218   of 2.5D axisymmetric simulations allowing exact representations of Earth's sphericity and of 3D

219   point sources. Our simulations have a maximum range of 2,000 km and frequency of 5 Hz,

220   typical values used in observations. We view our synthetic results as building on those of Yang

221   (2002) and Yang et al. (2007) who investigated *Lg* and *Sn* geometrical spreading for simple 1D

222   models, and so we do not include effects such as intrinsic attenuation or random scatterers. The

223   only factor that should make our synthetic results deviate from the 1D studies is the laterally

224   varying crustal thickness, which is also typically well-known, thereby allowing our results to be

225   quickly adapted to multiple regions of the world. We restrict our structural models to a Moho

226   ramp leading to crustal thickening. Crustal thickening has been less explored, perhaps due to its

227   subtler influence compared to crustal thinning, but this limited scope allows us to discuss

228   comparisons with real data (Fig. 2).

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230   After discussing the computational setup of our model, we explore individual *Sn* and *Lg*

231   amplitudes, and their amplitude ratios (*Sn/Lg*) in a 1D reference model (Figs. 4&5) and in Moho-

232   thickening models (Figs. 6–9), in which we establish the effectiveness of using *Sn/Lg* to identify

233   continental mantle earthquakes in the presence of significant Moho thickening. We next examine

234   real data from Tibet (Figs. 10&11) by directly comparing ramp-crossing (S events, Table 1&

235   Fig. 2) and non-ramp crossing events (non-S events, Table 1 & Fig. 2), and show that *Sn/Lg* is a

236   valid criterion for separating mantle from crustal earthquakes for the ramp crossing events just as

237   for the non-ramp-crossing events. Hence the *Sn/Lg* method, if used rigorously with local shallow

238   comparison events and multiple recording stations, can recognize the signature of a mantle

239   earthquake even with stations in a region of crust much thicker (Tibet) than the source region

240   (Indian Shield). Although we do not reliably detect effects of the ramp on individual *Sn* and *Lg*

241   amplitudes, we are able to confirm enhancement of high-frequency *Lg* across the ramp (Fig. 12)

242   due to neighboring mode-coupling during *Sn*-to-*Lg* conversion (Figs. 3, 13). Indeed, *Lg*

243   frequency content is another powerful discriminant for continental mantle earthquakes.

2. Computational aspects

245   We use the AxiSEM3D software package, whose main advances compared to previous 2.5D

246   axisymmetric methods (Bottero et al., 2016; van Driel et al., 2015) are that it can account for

247   fully 3D variations in terms of volumetric perturbations (Leng et al., 2016) as well as through

248   undulating surfaces (i.e. structural variations to either internal surfaces such as a Moho ramp or

249   external surface such as the ellipticity of the earth or topography) that break the spherical

250   geometry necessary for an axisymmetric method (Leng et al., 2019). We first briefly discuss

251   these new features from a user's perspective and introduce two necessary modifications made to

252   the source code in order to enable simulations with our desired scale and output.

253

2.1 Computational method and its enhancements

255   Without considering undulating surfaces, the azimuthal component of the 3D wavefield (from 0

256   to 2π in the plane perpendicular to the source-receiver direction) can be conveniently

257   represented by a Fourier series, which localizes the equations to a single meridian plane not

258   associated with any physical location, and then can be solved with a 2D spectral-element method

259   (Leng et al., 2016). Recognizing that lateral heterogeneities in earth are much smaller than

260   vertical ones, this hybrid scheme essentially uses "one line-shaped element" and high-order

261   Fourier series in the azimuthal direction, and in the 2D meridian plane uses 4th-order Lagrange

262   polynomials on a mesh with the quad-shaped elements that are necessary for a conventional

263   spectral-element method. The cost of 3D simulations in AxiSEM3D depends not on the length

264   of the 3rd dimension which in AxiSEM3D is always 0 to 2π, but on the 2D model size and

265   highest wave frequency, since these determine the number of elements that each have an

266   associated Fourier series. Even though the AxiSEM3D hybrid scheme is much more efficient for

267   a global 3D model than a fully 3D scheme, we note that a small 3D model needs the same

268   Fourier orders as the global model with the same level of lateral heterogeneity, and so a

269   conventional 3D method might be more desirable in this case. Testing for this specific threshold

270   is beyond the scope of this study. We estimated the Fourier orders needed if we were to extend

271   our 2.5D simulation to 3D based on Equation (5) in Szenicer et al. (2020), and found 3D

272   simulations are well out of reach given our available computational resources.

273

274   Even for our 2.5D simulations, the spatially and temporally down-sampled wavefield (discussed

275   in detail in the next section) on one meridian plane on one wavefield component is about 1.5

276   terabytes, and AxiSEM3D by default directs all parallel processes (MPI ranks) to output the data

277   to the same location. For our output size, the bandwidth to one location in a filesystem is

278   overloaded, greatly reducing the performance (since MPI ranks spend most of their time waiting

279   for I/O instead of computing) and more importantly, causing frequent filesystem crashes. We

280   take advantage of local hard drives physically connected to each computing node on Stanford

281   University's Sherlock HPC cluster (https://www.sherlock.stanford.edu/) through infiniband, and

282   we modified AxiSEM3D so that each MPI rank can identify its own computing node at runtime

283   and output its results to that node's physically-connected hard drive. This resolved the problem

284   of crashing the filesystem, and increased the performance of AxiSEM3D by at least one order of

285   magnitude for our problem size.

286

287   AxiSEM3D accounts for structural boundary variations diffeomorphic to a spherical or flat (for

288   Cartesian mesh) boundary through the use of a "particle-relabeling transformation" (Leng et al.

289   2019; Al-Attar &Crawford, 2016), which finds the change in radial coordinate for each

290   collocation point inside an element necessary to represent the structural boundary (and a finite

291   thickness transition zone around it). The input mesh is a 2D mesh without any geometric

292   variation (undulations on structural boundaries) or volumetric variation (perturbations of material

293   properties such as density and elastic constants), so is a 1D vertically-layered model compatible

294   with the axisymmetric requirement. The 2D and 3D variations are added on later with separate

295   files (for geometric and volumetric variations) and then described as Fourier coefficients for each

296   element in the 2D mesh. The 2D mesh can be related to a specific physical location and its

297   properties only by association with a specific azimuthal angle ϕ. The current version of

298   AxiSEM3D (Leng et al., 2019) does not output any of the built 2D or 3D models, and computed

299   wavefields can only be plotted on the coordinates of the spherical mesh with no undulating

300   surfaces. This causes a distortion of the wavefield visualizations that is too small to see on a

301   global scale (Nissen-Meyer, pers. comm.), but unacceptable for our regional-scale lithospheric

302 simulations. Further, although users can define models for undulating surfaces, there is currently

303 no way to check if this is being represented accurately inside the program. We modified the

304 source code to output the Fourier coefficients related to structural boundary variations for

305 affected elements, and then deform the input mesh to obtain 2D variations at any azimuthal angle

306 ϕ (constant in our 2.5D simulation) specified in our input geometric model. This enables us to

307 visualize our regional wavefield without distortion and to confirm that our Moho undulation is

308 exactly represented by the Fourier series (Supplementary material S1).

309

310 Lastly, we note that it is still not a straightforward and cheap task to perform regional wave

311 simulations even with the proliferation of computing resources and advancements in efficient

312 computational methods. Each of 51 simulations presented here required ~2 days with 500 cores

313 on Stanford's Sherlock supercomputer. Since most regional wave applications focus on

314 amplitudes, a much cheaper method based on radiative transport theory (a form of advanced ray

315 theory), that can only calculate absolute amplitude but can easily account for 3D structures and

316 random scatters (Sanborn et al., 2017) is potentially attractive. We did not use radiative transport

317 because we prefer a fully-numeric method, and because radiative transport has been shown to

318 underestimate shear energy both in 2D (Pryzbilla et al., 2006) and in 3D (Pryzbilla et al., 2008).

319

2.2 Model design

321 Our simulation domain is shown in Fig. 1. It has an effective size of ~2,000 x 230 km within the

322 absorbing boundary conditions. The vertical properties are based on PREM (Dziewonski and

323 Anderson, 1981) with the Moho depth adjusted to 30 km to better represent continental areas.

324 Our 230-km depth includes 10 km of the positive velocity gradient below the mantle low

325 velocity zone (LVZ) in order to include the LVZ trapped waves as well as Sa: the shear wave

326 trapped between the free-surface and the bottom of LVZ (Schwab et al., 1974; Wang and

327 Klemperer, 2023) that are important for the energy partitioning of surface waves.

328

329 Our reference model has Moho depth fixed at 30 km. For our other simulations, we introduce a

330 Moho ramp, in the shape of a sigmoid function at distance d from the source and with width w

331 that smoothly transitions the Moho depth from 30 km on the source side to 60 km beyond the

332 ramp (Fig. 1). The distance-to-ramp parameter d controls the portion of the wavefield that will

333   interact with the ramp and, with our fixed ramp height of 30 km, the ramp width parameter w

334   controls the steepness of the ramp. We vary d from 100 to 700 km with a 200 km interval, and

335   w from 100 to 400 km with a 100 km interval, values chosen to capture a wide range of realistic

336   scenarios. Our steepest ramp, that thickens by 30 km over a distance of 100 km ($>16°$ dip), is

337   analogous to the steepest part of the Himalayan Moho ramp, typically 20-25 km vertical change

338   over a 100-km width (e.g. Gao et al., 2016; Nabelek et al., 2009; Shi et al., 2016). The lowest

339   slopes we model, ~4° dip (here, a 30-km ramp spanning 400 km), is more characteristic of

340   eroded mountain belts in which total Moho relief of ~15 km within ~200 km across strike is

341   typical (e.g. Cook et al., 2010). However, it is commonplace for earthquakes to be recorded

342   along raypaths that are oblique, not perpendicular to orogens, and the first-order influence of

343   oblique incidence can be approximated through an increase of ramp width (Bostock & Kennett,

344   1990). Our gentlest ramp (w = 400 km) is analogous to that seen by an earthquake recorded at

345   45° obliquity to the Himalayan ramp. We use as our source a thrust earthquake with moment

346   magnitude $M_w = 6$, dip $\delta = 45°$, rake $\lambda = 90°$, and a Gaussian source time function with a half-

347   width of 0.2s. For each combination of ramp parameters w and d, as well as for the reference

348   model, we calculate the SH wavefield due to placement of this source at 3 different depths:

349   15 km (mid-crust), 35 km (shallow-lid, just below the Moho at the source but within the depth

350   range of the ramp), and 65 km (deeper-lid, below the Moho everywhere in the model).

351

352   We constructed our 2D finite-element mesh to balance accuracy and efficiency. Our absorbing

353   boundary conditions, including the thickness of the sponge layers, are set such that 97% of

354   reflections for waves with >0.5 Hz frequency are eliminated (Haindl et al., 2020). We use two

355   elements (10 collocation points) per wavelength in our simulations, and confirmed there was no

356   visible numerical dispersion. Using just less than two elements per wavelength – as needed for

357   the mesh coarsening – also produced no visible difference. We coarsened our mesh using two-

358   refinement transition templates (Anderson et al., 2009) (by "tricking" the built-in mesher) at a

359   depth of 70 km, 10 km below the deepest Moho, to ensure that mesh coarsening and deformation

360   (due to the Moho ramp) do not conflict. Our coarsening strategy enforced two elements per

361   wavelength at the coarsening depth, and more (or slightly fewer) elements per wavelength above

362   (or below) this depth, which resulted in a ~16.8% reduction in the number of elements needed

363   (Supplementary materials S2).

3. Numerical results from the reference model

As a check of our computational setup, and to ensure that *Sn* and *Lg* amplitude variations we discuss later are due only to the presence of the Moho ramp, we first calculate wavefields in a reference model with constant crustal thickness (no ramp) and compare our results to geometrical-spreading results calculated with a full-waveform method for a similar but non-identical 1D earth model for both *Lg* (Yang, 2002) and *Sn* (Yang et al., 2007) (Fig. 4). Both our source depths and distance range are larger than those explored by previous studies, so some discrepancy is expected apart from differences in earth models. We measure the *Sn* and *Lg* amplitudes at each offset as the RMS value over a time window defined by the expected range of group velocities, 4.0 to 4.7 km/s for *Sn* and 3.0 to 3.8 km/s for *Lg*. Our *Sn* windows are picked slightly differently compared to Wang and Klemperer (2021) to minimize overlaps with *Lg* windows at short distances and the mis-categorization of fast *Lg* waves at long distances, and also to account for non-zero source depths while not pre-judging whether an earthquake has a mantle or crustal hypocenter (Supplementary materials S3). All amplitudes are reported as displacements.

Fig. 4a&b show synthetic absolute *Lg* and *Sn* amplitudes, respectively, for the three source depths (15, 35 and 65 km) in the reference model (Moho depth fixed at 30 km), filtered from 1–5 Hz for *Lg* and around 3 Hz (from $3/\sqrt{2}$ to $3\sqrt{2}$ Hz, following Yang et al., 2007) for *Sn* and plotted at 10-km intervals from 200 to 2,000 km epicentral distance. Our frequency filters are 8th-order Butterworth filters. We also plot the relative amplitude decay provided by geometrical spreading models. For *Lg*

$$G_{Lg}(r) = r^{-\gamma}$$

where r is the epicentral distance and $\gamma = 1$ is an empirical constant (Yang, 2002). Yang (2002) modelled a variety of parameters such as source depth (but only tested crustal sources, above the Moho), frequency content, and amplitude-measurement technique, and found γ remained close to 1. *Sn* geometrical spreading is more complicated due to its propagation path (whispering gallery or interference head waves; cf. Avants et al., 2011), and has been modeled with both frequency (f) and distance dependence:

394
$$G_{Sn}(r,f) = \frac{10^{n_3(f)}}{r_0}\left(\frac{r_0}{r}\right)^{n_1(f)\log(r_0/r)+n_2(f)}$$

395 where $r_0 = 1$ km and $n_i(f)$ are fixed parameters for a specific two-layer Earth model, calculated

396 by Yang et al. (2007) for a source at 15 km depth in a uniform 40-km thick crust. The frequency

397 dependence of *Sn* geometrical spreading is the reason why we show single-frequency *Sn* in Fig.

398 4b.

399

400 We first note the preferential *Lg* excitation by crustal sources (amplitudes for the crustal source

401 are an order of magnitude greater than for the mantle sources: Fig. 4a) and preferential *Sn*

402 excitation by mantle sources (amplitudes for the 15-km source are an order of magnitude less

403 than for the 65-km source: Fig. 4b). This is reflected by the amplitude ratios *Sn/Lg* for these three

404 source depths, where although they vary as a function of epicentral distance, at each location,

405 *Sn/Lg* for mantle earthquakes is always higher than *Sn/Lg* for crustal earthquakes (Fig. 5). In an

406 ideal 1D model (i.e. the reference model), *Sn/Lg* varies according to epicentral distances in

407 different ways for different source depths, but the common feature is the increase of *Sn/Lg* at

408 long distances, i.e. about 600-1300 km, as noted before (Wang and Klemperer, 2021). The

409 amount of increase is the highest for the 15-km event. For this same 15-km event, there is also a

410 rather large decrease of *Sn/Lg* at shorter distances, whereas the values remain relatively constant

411 for the 35- and 65-km events. The reasons behind this is due to an over-estimation of *Sn*

412 following conventional observation methods, to which we return below while discussing

413 individual *Sn* and *Lg* amplitudes.

414

415 The higher-amplitude models (15 and 35 km for *Lg*, 65 and 35 km for *Sn*) vary fairly smoothly

416 with distance (Fig. 4 a&b). The abrupt discontinuities present in the bottom traces of Fig. 4a (65-

417 km trace) and Fig. 4b (15-km trace) are artifacts due to mis-categorizations of the waves in our

418 windowing process. For example, Fig. 4c,d&e show the synthetic seismograms for the 65-km

419 source at 400, 410 and 460 km (triangles in Fig. 4a&b). For this deep source, we expect

420 essentially no *Lg* excitation. However, at 400 km, there are two prominent *Sn* peaks included in

421 the *Lg* window (Fig. 4c), which explains why measured *Lg* amplitudes are unexpectedly high for

422 distances from 200-400 km (Fig. 4a). Measured *Sn* amplitudes are correspondingly lower than

423 the total amplitude within the *Sn* phase but by a smaller proportional amount because the later *Sn*

424   peaks are lower amplitude than the first-arriving *Sn* peak. At 410 km (Fig. 4d), the second *Sn*

425   peak is no longer within the *Lg* window, resulting in a sharp drop of measured *Lg* amplitudes. At

426   460 km (Fig. 4e) the second *Sn* peak moves into the *Sn* window, resulting in a small proportional

427   increase in measured *Sn* amplitudes for the 65-km source at that distance (Fig. 4b). This artifact

428   (peaks moving in and out of a window) is present for all cases in Fig. 4, but is small for the

429   major phase from each source depth, e.g. the small sinusoidal oscillations in measured *Lg*

430   amplitudes for the 15- and 35-km sources. These small variations were also shown but not

431   explained in previous synthetic studies (Yang, 2002). Although it is important that we fully

432   understand our synthetics, these phenomena have no relevance for real data for which small-

433   scale scatterers will always act to smooth out the strong amplitude peaks seen in our synthetics

434   (and those of Yang, 2002 and Yang et al., 2007). Our observations offer insight into the

435   relationship between physical *Sn* and *Lg* (as defined by propagation waveguides) and

436   observational *Sn* and *Lg* (as defined by group-velocity windows), and into the fundamental

437   inaccuracy of using the same *Sn* and *Lg* velocity windows for different events because these

438   window-bounding velocities are dependent on epicentral distance even for the same source depth

439   in a 1D model (a point also touched on by Aki and Richards, 2002, their Box 7.1). For example,

440   all three wavelets in Fig. 4c are physical *Sn* waves, ordered by their relative mantle and crustal

441   path lengths, and the *Sn* velocity window would need to be extended down to 3.4 km/s to capture

442   all three in the *Sn* window. At just 400-km range, the second and third wavelet appear in the *Lg*

443   window (and also a small wavelet around 127s, Fig. 4c), but with increasing offset all these

444   peaks would have travelled proportionally greater distances in the mantle, gaining higher

445   apparent horizontal group velocity, and be captured in the conventional *Sn* velocity window (Fig.

446   4d&e).

447

448   Working with synthetics it would be possible to measure the apparent group velocity of each

449   arrival and thereby correctly separate *Sn* from *Lg*; but in real data such an approach is likely

450   difficult or impossible. Hence we do not seek to change the conventional observation method,

451   but rather we acknowledge the prevalence of this issue and highlight the irrelevance of fine-

452   tuning velocity windows and the care needed to avoid over-interpreting amplitude

453   measurements. Fortunately, the conventional and tractable *Sn*- and *Lg*-velocity windows method

454   are historically proven to be adequate, especially when only crustal sources are considered. Since

our *Sn/Lg* method for identifying mantle earthquakes fundamentally relies on comparisons between potentially crustal and mantle earthquakes, rather than absolute-value *Sn/Lg* thresholds, it is more important to use a simple and unified approach for a group of events (to enable comparisons) than to strive for picking the most accurate windows for individual events, which in practice is also hard to achieve.

Sn synthetic amplitudes (Fig. 4b) exhibit the classic interference head-wave behavior at distances <~1,300 km, in that amplitudes first decrease then increase with distance due to the spherical focusing effect, because at larger distances more energy from waves multiply-reflected at the Moho underside will contribute to the amplitudes, in addition to the direct arrival. The distance at which the *Sn* amplitudes begin to increase and the amount of the increase depends on source depth, and is closest/strongest for the shallowest source (note this is not captured by the geometrical spreading model, black lines in Fig. 4b, as that is an empirical fit based on a crustal source only). For the 15-km source, we expect mostly *Lg* excitation. While our *Sn* window is already shortened compared to Wang and Klemperer (2021), the earliest *Lg* waves could appear within the *Sn* window, which results in the artificial amplitude jump at 820- and 830-km distances (circles in Fig. 4b; Fig. 4f&g). As before, these sudden amplitude changes are due to mis-categorizations, but are likely much smaller in real data due to presence of smoothing effects. However, if these smoothing effects are not accounted for and if the conventional windowing method is followed (Yang et al., 2007) (completely justified if the intent is to study geometrical spreading alone), the amount of *Sn* amplitude increase might be over-estimated at these long offsets due to incorporation of *Lg* waves. This incorporation of *Lg* waves at long offsets potentially explains the earlier rise to larger *Sn* amplitudes, leading to a larger increase of *Sn/Lg*, for the crustal source compared with the mantle sources (Fig. 4b, Fig. 5). For a 35-km source, the smooth amplitude increases from e.g. 900- to 1,000-km distances (squares in Fig. 4b) are due to the spherical focusing effect of interference head waves, as the number of peaks within the *Sn* window is not changed, yet their amplitudes (most notably the third peak) grow larger (Fig. 4h, i). For distances >1,300 km, all three sources have about the same amount of *Sn* energy (Fig. 4b), but their *Lg* energy is vastly different (Fig. 4a) so our method is still very effective at these long offsets. Beyond ~1,300 km our measured amplitudes start to drop, a phenomenon not previously noted because the Yang et al. (2007) study was limited to shorter offsets, but completely reasonable because the spherical focusing effect must eventually wear off, i.e. the multiply-reflected waves at the Moho underside

486 eventually become too small to meaningfully contribute. In all other respects, our 3-Hz *Sn*
487 amplitudes exhibit the same interference head-wave behavior as in Yang et al. (2007). In fact, the
488 fit to our 15-km source (the same source depth simulated by Yang et al. (2007), but in a slightly
489 different earth model) is good (Fig. 4b). The misfit to the deeper sources beyond ~500 km clearly
490 originates from the fact that only a crustal source was considered by Yang et al. (2007), which,
491 combined with other reasons discussed above, led them to an over-estimation of the spherical
492 focusing effect.

493

494 We have an almost exact match between our 15- and 35-km sources' synthetic *Lg* amplitudes
495 and the simple *Lg* geometrical spreading model (Yang, 2002), and between our 15-km source's
496 *Sn* synthetic amplitudes and the more complicated *Sn* geometrical spreading model (Yang et al.,
497 2007). The mismatches between our synthetic amplitudes and the previous models can all be
498 understood. This gives us confidence in our modelling approach. Our results cover a larger
499 parameter space and exhibit a greater range of features than previous studies, so already provide
500 useful new information as well as serving as a benchmark against which to test our simulations
501 with a Moho ramp.

502 ## 4. Numerical results from Moho ramp models

503 Our parameter-space study includes 48 2.5D crustal-thickening simulations (plus reference
504 simulations). Here, we present a selection of these results (Figs. 6–9) as two groups by first
505 fixing the ramp width $w = 200\ km$ and varying the distance to the start of the ramp $d =$100,
506 300, 500 and 700 km; and then by fixing $d = 300\ km$ but varying $w =$ 100 km ~17°, 200 km
507 ~9°, 300 km ~6° and 400 km ~4°. The rest of our numerical results can be found in
508 Supplementary materials S4&5. The amplitudes are measured as discussed in Section 3, except
509 that now our *Sn* amplitudes are measured using the same broader frequency band we use for *Lg*,
510 i.e. 1-5 Hz. We present our results for *Sn* and *Lg* amplitudes (Figs. 6, 9, left and middle
511 columns) as ratios to our reference-model results, aligned by distance relative to the end of the
512 ramps to highlight deviations relative to ramp locations. We show results for *Sn/Lg* (Figs. 6, 9,
513 right columns) relative to epicentral distance and overlaid on reference-model results to highlight
514 the ramp effects and to illustrate the absolute values of *Sn/Lg* for different source depths.

515

516   Our interest is in phenomena that have the potential to be recognized and measured in real data.

517   Measurements on synthetics of *Sn* alone, or *Lg* alone, coupled with inspection of synthetic

518   seismograms and compared to the reference (flat Moho) model reveal the physics of wave

519   propagation across Moho ramps. However, in the real world no reference data are available and

520   it is the *Sn/Lg* ratios that, by removing source and receiver dependencies, may allow recognition

521   of source depth with respect to Moho (Wang & Klemperer, 2021).

522

523   ## 4.1 Fixed ramp width, $w = 200\ km$

524   This fixed ramp width, combined with our constant 30-km ramp height, produces a Moho ramp

525   with fixed gradient that is among the steepest in nature (though not the steepest Moho ramp we

526   test, see Supplementary materials S4&5), especially considering that many earthquake-receiver

527   geometries involve oblique incidence onto the ramp, effectively increasing the width of the

528   ramp. We highlight this example, $w = 200$ km, because incidence onto a steep Moho ramp

529   produces the clearest effect on *Sn* and *Lg* amplitudes. For all our simulations (variable w, d and

530   source depth z) the measured amplitudes (and the *Sn/Lg* ratios) coincide with those from the

531   reference model until the waves reach the ramp apart from tiny numerical errors (e.g. Figs. 6a,d

532   show some symbols at distances -700–0 km slightly below the grey line that represents equality

533   with reference model results) .

534

535   *4.1.1 Sn amplitudes*

536   For relative *Sn* amplitudes (Fig. 6, left column, a.-c.), one of the most striking features is the

537   focusing of *Sn* waves that starts close to the middle of the ramp and peaks slightly beyond the

538   ramp, for most source depths z and distances to ramp d. This phenomenon corresponds to the

539   breaking of the *Sn* waveguide by the Moho ramp (Fig. 7). Before the leading wavefront in the

540   mantle reaches the ramp (at time = 73 s, Fig. 7a), the wavefield is the same as in the reference

541   model, with the same first reflected wave as the leading wavefront and the same first arrival at

542   the surface (Fig. 7a&b). At time = 100s the leading wavefront is at about the middle of the ramp

543   and the first arrival is just beyond the start of the ramp (Fig. 7c). At 100s, the Moho underside

544   transmitted wave has a much larger amplitude in the ramp model than in the reference model

545   (Fig. 7d), and this increased amplitude extends to the surface, representing the start of the *Sn*

546   peak just beyond the start of the ramp. The reasons behind the focusing are twofold. First, the

547   Moho ramp increases the local curvature of the Moho so that deeper energy on the leading

548   wavefront, which in the reference model would refract up at a greater distance (Fig. 7e, blue

549   arrows), refracts up to the surface from the ramp (Fig. 7e, yellow dashed arrows), locally

550   increasing the amount of energy being transmitted into the crust. Second, the incidence angle of

551   this deeper energy changes from almost grazing to a smaller angle ($i_{ramp} < i_{flat}$) (Fig. 7e),

552   which could flip the energy partitioning of the reflected and transmitted waves (Fig. 7f) in favor

553   of transmission (calculated using plane-wave transmission and reflection coefficients, e.g. von

554   Seggern, 2012, which are a good approximation at long distances from the source). The large

555   reflected energy at large incidence angles (Fig. 7f) enables multiple reflections at the Moho

556   underside, and essentially gives rise to the whispering-gallery waveguide. The increase in

557   transmitted energy at decreased incidence angles shows how this waveguide is broken by a

558   Moho ramp. On a seismogram (Fig. 7g), other than the prominently increased amplitude of the

559   first arrival of the ramp model, the effect of the ramp shows up as delays in individual arrivals

560   due to the transmitted waves travelling a longer distance in the crust and travelling at a steeper

561   angle ($r_{ramp} < r_{flat}$, Fig. 7e) leading to a smaller horizontal apparent velocity. In general, the

562   magnitude of this focusing (~2 to 10 times stronger than reference, Fig. 6, left column) is a proxy

563   for how much of the original wavefield interacts with the ramp, which is inversely proportional

564   to d and is largest for the 35-km event (that lies within the vertical extent of the ramp), followed

565   by the 15- and 65-km events. This explains why the 65-km deeper-lid event has about half the

566   focusing strength of the shallow-lid earthquake (35-km) and the mid-crustal earthquake (15-km),

567   which both have similar degrees of focusing. For the 35-km event, the focusing strength is

568   strictly inversely proportional to d, but this is not the case for the 15- and 65-km events for

569   which, for shorter distances to the ramp, more complicated interferences occur that decrease the

570   strength of the Sn peak ($d = 100$ for 15-km event, Fig. 6a, and $d = 100, 300$ km for 65-km

571   event, Fig. 6c).

572

573   The only case without an Sn peak at ramp exit is z=15 km, d=100 km (Fig. 6a). Regardless of

574   ramp width, no crustal sources at this short distance to ramp show an Sn peak, but instead have a

575   large decrease of Sn for longer ranges beyond the ramp exit, and all eventually recover back to

576   close to reference values (Fig. S5-1a). Similarly, Sn de-focusing is present for the 65-km events

577   when d is short (i.e. 100 and 300 km) (Fig. S5-1c). Unlike the Sn focusing peak that is just

578   outside (<200 km) of the ramp exit (left column of Figs. 6, S4-1), or even completely contained

579   inside the ramp region for wider ramps (left column of Figs. S4-2, S4-3), these Sn de-focusing

580   regions can extend up to 600 km (Fig. 6a) to 800 km (Fig. 6c), having a broad influence on Sn/Lg

581   (e.g. Fig. 6i, $d=100$ &300 km). It would seem that e.g. if an earthquake with $z = 65$ km, $d = 100$

582   km is measured at ~700-800 km epicentral distance (Fig. 6i), its Sn/Lg could be confused with

583   that of an earthquake with $z = 15$ km and $d = 500$ km measured at the same distance (Fig. 6g).

584   However, such a confusion requires a careful orchestration of a broad de-focusing zone and a

585   localized Sn focusing peak, as well as potentially very different back-azimuths to a particular

586   station to produce the 400-km difference in effective ramp width, and although this could occur

587   in the real data, array-based measurements with varying source-station geometry should be able

588   to mitigate, if not completely avoid, this effect.

589

590   The behavior of Sn is quite complex in the presence of a crustal-thickening Moho ramp due to its

591   interference head-wave nature. For example, there are also secondary focusing peaks for the 35-

592   km events for all d's (Fig. 6b). However, such specific observations on synthetics are likely too

593   detailed to observe in real data, so we do not further discuss or decipher these phenomena.

594

595   Lastly, in contrast to the above-mentioned deviations from the reference model, the other most

596   striking feature, common to all our simulations (Figs. 6, 9, left columns, Supplementary

597   materials, S4&5) is that, at long offsets, Sn amplitude always returns to about the same level as

598   the reference model. The mechanism for unperturbed amplitude at long offset is that deeper

599   energy (below the black dot on the wavefront, Fig. 7e) may never interact with the ramp, and

600   hence at long distances this energy is transmitted into the crust as if the Moho had always been

601   flat beneath a thickened 65-km crust. However, arrival delays (Fig. 7g) persist for the ramp

602   model even at long distances beyond the ramp, simply due to the increased travel path in the

603   crust.

604

605   *4.1.2 Lg amplitudes*

606   Clear Sn-to-Lg conversion due to Moho thickening can be seen in Fig. 6, middle column,

607   especially for the mantle sources. To first order, when the ramp shape is fixed as in the present

608 case (i.e. height = 30 km and w = 200 km), we expect the degree of conversion to be controlled

609 by the amount of *Sn* excited (positively correlates with source depth z) and the subset of this

610 amount that interacts with the ramp (negatively correlates with d and source depth relative to

611 Moho). Our results indicate that the source depth plays a far more important role. For the 65-km

612 event, all d's share a similar growth pattern for relative *Lg* amplitude, which increases by >5

613 times between the ramp start and end, due to *Sn*-to-*Lg* conversion. At longer offsets, beyond the

614 ramp, the continued gradual increase in relative *Lg* is due to the decrease in *Lg* for the reference

615 model for this sub-crustal source (Fig. 4a, magenta triangles) rather than to *Lg* growth in the

616 ramp models. For the 35-km event, even though the source depth is within the vertical extent of

617 the ramp, *Sn*-to-*Lg* conversion is relatively modest (Fig. 6e). Except for the source closest to the

618 ramp ($d = 100km$), the relative *Lg* amplitude increase from the reference model is less than a

619 factor of 2, which means real-world observations perturbed by scatterers and noise could be

620 difficult. Beyond the ramp, after the initial oscillations in relative *Lg* amplitudes, we see relative

621 amplitudes decrease. This is especially prominent for $d = 300$ and 500 km, and is subtle for $d = $

622 100 km perhaps due to more initial wavefield interaction with the ramp, and is not shown for

623 $d = 700km$ because our 2000-km maximum simulation range does not include distances

624 sufficiently far beyond the ramp (Fig. 6e). As ramp width increases, drop-off of amplified *Lg*

625 becomes even more pronounced (e.g. Figs. S4-2e, S4-3e). The crustal 15-km source (Fig. 6d)

626 does not show this same behavior of relative-amplitude decrease, even though in the reference

627 model *Lg* decays at the same rate for both 15-km and 35-km sources. The relative-amplitude

628 drops for the 35-km below-Moho source may therefore indicate that the crustal waveguide

629 cannot sustain the increased *Lg* frequencies that are created by *Sn*-to-*Lg* conversion, a topic we

630 return to below (see Fig. 8).

631

632 Relative *Lg* amplitude from the 15-km source decreases by factor <2 as we cross the ramp (Fig.

633 6d). The magnitude of decrease is inversely proportional to ramp distance d because disruption

634 of the *Lg* waveguide (i.e. the crust) causes *Lg* de-focusing as illustrated by Kennett (1986).

635 Though this de-focusing must also occur for the 35-km and 65-km sources it is more than

636 compensated for by the strong *Sn*-to-*Lg* conversion from these deeper sources. For the 15-km

637 mid-crustal source, *Sn*-to-*Lg* conversion is hard to observe as there is much less initial *Sn* energy

638 that can be potentially converted to *Lg* (Fig. 4b). However, *Sn*-to-*Lg* conversion must still be

639 occurring because at larger distances beyond from the ramp relative Lg amplitude for all d's

640 increases above 1, indicating extra Lg energy than expected if the Moho was uniform.

641

642 When Sn converts to Lg we expect not only amplitude but also frequency effects: the higher-

643 frequency portion of the Lg becomes enriched because mode coupling tends to happen at

644 neighboring modes (Fig. 3) (Maupin et al., 1989), thus higher-mode Sn tends to excite higher Lg-

645 forming modes, which contribute to Lg at higher frequencies. We test this with our full-

646 waveform results by comparing the Lg wavetrain filtered from 1–5 Hz, as shown thus far, to the

647 same wavetrain filtered 0.1–0.8 Hz. We plot the high-frequency (HF) to low-frequency (LF)

648 ratio (Lg HF/LF) (Fig. 8) of the ramp model divided by the reference model, and confirm that Lg

649 has a higher frequency component that develops across the ramp due to Sn-to-Lg conversion. For

650 the 15-km source, although subtle, there is a slight increase of high-frequency content further

651 away from the ramp, which suggests some Sn-to-Lg conversion for the mid-crustal source.

652 Comparing Fig. 8b&c to Fig. 6e&f we see similar trends, implying that a large part of the

653 increase in Lg beyond the ramp is due to the increased HF component from Sn-to-Lg conversion.

654

655 *4.1.3 Sn/Lg amplitude ratios*

656 The changes in Sn or Lg amplitude or frequency content, relative to the reference model, are by a

657 factor typically <2, so can be hard to recognize on real, noisy, data (except for the deepest source

658 at the largest offset, Figs. 6f, 9c). In contrast, the amplitude ratio Sn/Lg is a direct measure of the

659 relative strengths of Sn and Lg amplitude perturbations. Sn/Lg in the reference model increases

660 linearly in log amplitude–log distance space for distances from ~600–1,400 km (Figs. 5, 6&9 g-

661 i), confirming our earlier conclusion (Wang and Klemperer, 2021) from analysis of the empirical

662 geometrical spreading models that differ between Sn and Lg. When the source is at 15 km, Sn/Lg

663 largely follows the shape of Sn variations (Fig. 6g), because the Sn-to-Lg conversion is rather

664 weak. Since any ramp only locally perturbs Sn amplitude perturbations, if amplitude ratios are

665 measured far enough beyond the ramp, there is virtually no difference between the ramp and the

666 reference models. Hence, crossing a significant Moho ramp (as in the present example) does not

667 affect Sn/Lg observations for a crustal earthquake provided the measurements are made

668 sufficiently far beyond the ramp. Exactly how far is sufficient is related to d, and ranges from

669 ~800 km beyond the ramp for $d = 100\ km$ (measured from the red vertical line in Fig.6g) to

670   ~100 km beyond the ramp for $d = 700\ km$ (measured from the cyan vertical line in Fig.6g).

671   More simply, for source depth = 15 km for all d studied here (100 km $\leq d \leq$ 700 km), a

672   propagation distance of ~1100 km is sufficient to erase most of the ramp effect on *Sn/Lg*: at this

673   distance all symbols coincide with the reference model (grey inverted triangles, Fig. 6g). Lastly,

674   we note that *Sn/Lg* almost nowhere exceeds 0.2 for the 15-km event, marked by a black fiducial

675   line in Fig. 6g,h,i.

676

677   In contrast, for the 35-km and 65-km sources, *Sn/Lg* is typically an order-of-magnitude larger

678   than for the crustal source and rarely drops below 0.2 regardless of their variations, except

679   sometimes just beyond the ramp. For the 35-km source (Fig. 6h) we see a combined effect from

680   *Sn* and *Lg* variations, with *Sn* controlled by the local, transient focusing behavior, primarily in

681   the ramp region, and *Sn/Lg* determined largely by *Lg* amplitudes further beyond the ramp. The

682   biggest decrease of *Sn/Lg*, to ~3 times lower than the reference model at epicentral distance ~450

683   km is for the source that is closest to the ramp start and is due to increased *Lg* amplitudes (Fig.

684   6h, red circles). For the 65-km source, *Sn/Lg* is primarily controlled by the amplitude variations

685   of *Lg*, and in some cases with small d, *Sn* de-focusing. Further away from the ramp there are

686   significant decreases from the reference model by more than an order of magnitude, with the

687   source closest to the ramp again exhibiting the largest decrease. Even though *Sn/Lg* for the two

688   sub-Moho sources (Figs. 6h,i) far beyond the ramp can be smaller than had there been no Moho

689   ramp, *Sn/Lg* remains 5-10 times larger than for the mid-crustal, 15-km, event. Visually, crustal

690   and mantle earthquakes can be largely separated by the black fiducial line (Fig. 6g,h,i). In

691   consequence, the *Sn/Lg* method (Wang & Klemperer, 2021) is robust for all cases tested here,

692   especially if the recording stations are not limited to the ramp region and some measurements are

693   made far beyond the end of the ramp.

694

695   **4.2 Fixed distance to ramp, $d = 300\ km$**

696   Here, d is fixed, so all ramps start at the same location but they extend out by different distances

697   from $w = $ 100, 200, 300 and 400 km (Fig. 9). Since our ramp height is fixed, by increasing w,

698   we are decreasing the steepness of the ramp. As the dip of the Moho ramp decreases, the

699   amplitudes will converge to the reference model.

700

701   Many key observations remain the same as in the previous section, including the *Sn* focusing
702   peaks being localized close to the ramp and then recovering to the reference model at long
703   distances (Fig. 9, left column), and the increase of the *Lg* amplitudes (quite subtle for the 15-km
704   mid-crustal event) beyond the ramp that persists to greater distances in the crustal waveguide
705   (Fig. 9, middle column). Most important, even though the *Sn/Lg* ratio for each source depth
706   varies by a factor of ~20 with offset (Fig. 9, right column), *Sn/Lg* for the 35- and 65-km events
707   exceeds *Sn/Lg* for the 15-km event at all distances, typically by a factor of 5–10. Thus – as seen
708   also from Section 4.1 and Fig. 6, right column – *Sn/Lg* ratios are a robust metric for interpreting
709   source depth above or below the Moho even in the presence of a Moho ramp, but particularly
710   beyond the end of the ramp (Fig. 9, right column).
711
712   Increases in steepness of the Moho ramp can be thought of as an effective increase in the local
713   Moho curvature. The tighter the curvature, the more intense the upward focusing of *Sn* from the
714   Moho underside (Fig.7, e&f), leading to the most prominent feature in our simulations, the factor
715   of 2–10 increase in *Sn* compared to the reference model vertically above and immediately
716   beyond the ramp (Fig. 9a-c). Fig. 9a-c shows the steepest Moho ramp (*w*=100km) leads to the
717   strongest and earliest (with respect to epicentral distance) *Sn* amplitude increase, reaching peak
718   *Sn* amplitude just beyond the ramp region (for ramp widths <200 km). For wider ramps (*w* =
719   300 and 400 km), the peaks are completely within the ramp region. Just as for the models in Fig.
720   6, these results imply the need for observations across the ramp region if they are to be relevant
721   for real data. The *Lg* energy increase beyond the ramps is again well-aligned with the start of the
722   ramps (Fig.9, d-f) and is most prominent for the 65-km source, and decreases as the source
723   depths decreases. As expected, the gentlest ramp (*w*=400 km) has the smallest increase in *Lg*.
724   Ramp width seems to have less influence on *Sn* de-focusing. For the 15- and 65-km events, *Sn*
725   de-focusing occurs except for *w*=100 km when *z*=15 km (Fig. 9a,c), and for *d*=100 km, even
726   this exception doesn't hold anymore (Fig. S5-1a,c). However, at *d*=300 km, *Sn* de-focusing
727   barely influences *Sn/Lg* (Fig. 9g&i) and the de-focusing becomes completely absent for larger
728   *d*'s (Figs. S5-2, S5-3, left column).

5. Observational results

We study earthquakes recorded by the HiCLIMB array not only for its high data quality, but also because the Moho structure beneath this array is well-studied (Nabelek et al., 2009), providing good definition of the Moho ramp structure (Fig. 2). Along the HiCLIMB profile (IRIS data code XF), the Moho is relatively flat beneath northern India and the Main Frontal Thrust (MFT). The Moho ramp begins about 100 km further north, beneath the Main Central Thrust (MCT). Because the surface trace of the MCT is tortuous due to laterally varying exhumation of a low-angle structure (Martin, 2017), we use a line 100-km north of the MFT as our proxy for the start of the ramp (Fig. 2a). North of the MCT, the Moho deepens from ~45 km to ~65 km over a distance of 150 km to the Yarlung-Zangpo Suture (YZS), a geometry present all along the Himalayan arc (e.g. Gao et al., 2016; Shi et al., 2016). Our data set includes very few southern stations over the northern part of the ramp (usually ≤7 due to limited operating time and noisier data), with most available stations lying further north beyond the YZS. This station distribution offers only a glimpse close to the ramp region to investigate effects on individual Sn and Lg amplitudes but gives ample opportunity to observe Sn/Lg away from the end of the ramp, where synthetics predict its effectiveness. In addition to this main dataset, we also analyzed 4 events from the Gangdese-92 array (Fig. 2). These 4 earthquakes are directly due south of the stations, offering an opportunity to evaluate the influence of oblique incidence for the HiCLIMB events. We use the same Sn and Lg velocity windows as used in Wang & Klemperer (2021), based on regional observations in our study area, i.e. 4.3-4.8 km/s for Sn (McNamara et al., 1995) and 3.1-3.6 km/s for Lg (McNamara et al., 1996). The data is bandpass filtered from 1-5 Hz with an 8th order Butterworth filter. We select traces only if either or both Sn or Lg has a root mean square (RMS) amplitude at least twice as high as that of a noise window, defined to start 30s and end 5s before the Pn arrival that we calculate using a constant velocity of 8.1 km/s.

There have been reports of sub-crustal earthquakes beneath southern Nepal and northern India (e.g. Chen & Molnar, 1983; Chen & Yang, 2004; Baur, 2007; Song & Klemperer, 2024) but none are confirmed and counter-claims exist that all earthquakes in these regions are likely intra-crustal (Maggi et al., 2000; Mitra et al., 2005; Priestley et al., 2008). Here, we look at six earthquakes, our 'southern events' (S1–S6 from south to north) with catalog depths 10–62 km (Table 1), recorded on the HiCLIMB array (Nabelek et al., 2009; Fig. 2). The travel paths of S1–

760   S6 traverse a Moho ramp ~20 km high at distances $50 \leq d \leq 660$ km from the source, and

761   spanning widths measured obliquely along the path $160 \leq w \leq 475$ km. This ramp is smaller than

762   used for some of our synthetics, and the obliquity of raypaths to both the ramp and the HiCLIMB

763   profile means that different recording stations have a different d and w. This varied geometry is

764   beneficial to our method as it helps to avoid systematic errors. HiCLIMB also allows us to

765   compare our six southern events that do traverse the ramp with six earthquakes in northwestern

766   Tibet (Wang & Klemperer, 2021) (Fig. 2) that do not traverse a ramp or indeed any major Moho

767   topography. These six 'northern events', spanning upper-crustal to upper-mantle hypocentral

768   depths (Wang and Klemperer, 2021), were recorded on the same array as the southern events,

769   over roughly the same distance ranges (Supplementary materials S6). The southern and northern

770   events are also similar in magnitudes, ranging from m_b 3.5-4.3, with S4 being the smallest event

771   studied (Table 1). The HiCLIMB array operated only from mid-2004 to end-2005 and seismicity

772   in Northern India is not nearly as prolific as on northwestern Tibet, so our six southern events

773   have a much larger spatial spread than the six northern events.

774

775   We look at our data from three perspectives to illustrate the effect of regional waves traversing

776   through a Moho ramp: gross amplitude measurements (Fig. 10), individual seismogram changes

777   across a record section for a given event (Fig. 11), and Lg HF/LF, i.e. ratios of Lg amplitudes in

778   1-5Hz (HF) and 0.1-0.8Hz (LF) frequency ranges (Fig. 12).

779

780   **5.1 *Sn, Lg* amplitudes and *Sn/Lg* ratios**

781   We plot *Sn*, *Lg* and *Sn/Lg* for all our events against station distance north of the end of the ramp

782   (Fig. 10), YZS (Fig. 2), so that the horizontal axis is also a proxy for station locations allowing

783   evaluation of site effects along the array. We normalize the individual *Lg* and *Sn* amplitudes to

784   the first recording station, i.e. the southernmost and northernmost stations for southern and

785   northern events respectively, to remove first-order differences between earthquakes, e.g. their

786   different magnitudes. As we move north from the end of the ramp, epicentral distances increase

787   for the southern events, but they decrease for the northern events (Fig. 10). Alternatively,

788   plotting both groups of events against epicentral distance shows that amplitude generally

789   decreases as epicentral distance increases (Supplementary Fig. S6-1).

790

791 Individual *Sn* and *Lg* amplitudes for both southern (Fig. 10 a&b) and northern (Fig.10 c&d)

792 events show remarkable coherence as a function of station location, despite the many differences

793 within and between the two groups. At about 200-300 km beyond the end of the ramp, all

794 measurements (Fig. 10 a-d) are amplified. For the southern events (Fig. 10 a&b), this increase of

795 *Sn* and *Lg* amplitudes superficially resembles the *Sn* peak and increased *Lg* due to *Sn*-to-*Lg*

796 conversion predicted by our modelling (Fig. 6&9, left and middle columns). However, our

797 synthetics show both increases should occur closer to the end of the ramp, reaching their maxima

798 within 100–200 km beyond the end of the ramp. The distance of these maxima from the end of

799 the ramp decreases as ramp width increases, and the obliquity of our source-receiver azimuths to

800 the ramp creates very large effective ramp widths (Table 1). Hence the location of the *Sn* and *Lg*

801 maxima moves even closer to the ramp (Fig. 9, left & middle columns) so the amplitude

802 increases at 200–300 km (Fig. 10a&b) are most unlikely related to traversing the ramp. Indeed,

803 the northern events also show *Sn* and *Lg* amplification at the same stations, implying the peaks at

804 200–300 km are likely due to local variation in crustal and mantle seismic attenuation (Fig. 2).

805

806 Another potential candidate for an *Sn* focusing peak is shown by the few stations that recorded

807 the southern events within and closely adjacent to the ramp (~-50 – +100 km) (Fig. 10a). This is

808 promising because the northern events (Fig. 10c) do not seem to show this peak, and for our

809 southern events that traverse wider (less-steep) ramps the *Sn* peak should occur within the ramp

810 region (Fig. 9, left column). However, we are not confident that this is a true observation of an

811 *Sn* focusing peak because our secondary dataset from the Gangdese-92 array does not show the

812 same feature (Supplementary materials S7). The lack of an *Sn* focusing peak on the Gangdese-92

813 array, that recorded events with almost perpendicular incidence to the Moho ramp, implies that

814 obliquity of ray-paths to the ramp is likely not the cause of our inability to observe amplitude

815 variations due to the ramp. Observations of individual amplitudes in real data are subject to many

816 variables such as site effects (which likely is strong in the HiCLIMB data based on the coherence

817 seen in Fig.10 a-d), anelastic attenuation, and small-scale heterogeneities that could completely

818 erase the *Sn* and *Lg* ramp-traversal signatures in our synthetics.

819

820 The lack of unequivocal observations of ramp effects in the *Sn* and *Lg* amplitude data is

821 disappointing in that we cannot confirm the predictions of our synthetics from an amplitude

822   perspective, but the negligible influence of the ramp on amplitudes is a *positive* result for the

823   ability of the *Sn/Lg* method to distinguish below-Moho from above-Moho earthquakes. The

824   *Sn/Lg* method is robust because it is largely immune to site effects, due to ratioing of the two

825   portions of the same waveform recorded at the same location. Hence *Sn/Lg* ratios (Fig. 10e),

826   unlike individual *Sn* and *Lg* amplitudes (Fig. 10 a-d), do not show any strong correlation with

827   station locations and *Sn/Lg* ratios span similar values for both the southern and the northern

828   earthquakes. We can separate our events into two groups either visually (Fig. 10e), or more

829   quantitatively according to whether at least half of station *Sn/Lg* values are above or below our

830   previous experimental threshold for this region (Wang and Klemperer, 2021), *Sn/Lg* =2.

831   Southern event S1 has a single station and S6 has no station recording *Sn/Lg* >2 (Figs.10, 11): we

832   believe both are crustal earthquakes. In contrast, events S2, S4, and S5 have >50% stations

833   reporting *Sn/Lg* >2 (Figs.10, 11), and visually they behave like northern events WT1 and WT2

834   (Fig. 10e), which have previously been identified as upper-mantle events (Wang and Klemperer,

835   2021). This distinction is particularly clear >100 km north of the end of the ramp, and remains

836   clear across most of the northern attenuation zone. Measured across all the stations, southern

837   event S3 has just 39% of measurements with *Sn/Lg* > 2 (Fig. 11), but this rises to 52% if we only

838   consider stations >100 km north of the ramp end (Supplementary materials S8). If the catalog

839   depths for S3 and S4 are correct (~60 km) they are certainly below-Moho events. A full-

840   waveform inversion put S3 at 53 km (Baur, 2007), clearly below the local Moho (Singh et al.,

841   2015; Mitra et al., 2018), a conclusion (weakly) supported by our *Sn/Lg* results. S5 has an

842   arbitrarily assigned depth of 10 km, which is not a useful determinant of the real depth, and

843   based on the *Sn/Lg* data we believe it is in fact a sub-Moho event. Events S1 and S2 have depths

844   ~35 km, around Moho depth (Singh et al., 2015; Mitra et al., 2018) yet our method suggests S2

845   occurred below the Moho and S1 above it. S6, with a relatively reliable catalog depth of 16.1

846   km, in the upper crust, is also suggested by our *Sn/Lg* criterion to be a crustal earthquake. These

847   results show that although there is in general a positive correlation of *Sn/Lg* measurements with

848   catalog depth (Song and Klemperer, 2024), there could also be inconsistencies particularly for

849   the case of S5. Because comparison between different Himalayan catalogs shows numerous large

850   depth discrepancies (Song and Klemperer, 2024), and dedicated re-location efforts have found

851   some egregious catalog mis-locations (Craig et al., 2023), we suggest that our determination of

852 S5 as a sub-Moho earthquake from its *Sn/Lg* character may be more reliable than the assigned

853 catalog depth.

854

5.2 Record sections

856 To further investigate the excitation of *Sn* and *Lg* for the southern events, we turn to their record

857 sections (Fig. 11). normalized to the maximum value on each trace to highlight relative

858 amplitude changes within a trace. For our current dataset, the *Sn* and *Lg* windows do not overlap,

859 making their amplitude measurements distinct.

860

861 For the four events that we believe are of mantle origin (S2, S3, S4 and S5), clear *Sn* excitation

862 can be observed in the middle part of the record section, at distances >~100 km north of the YZS

863 (the Moho ramp end, labelled as 0 on the upper x-axes of the record sections, Fig. 11). At

864 distances >~400 km beyond YZS there is some diminution of *Sn*, as waves reaching these

865 stations have propagated partly within the region of high *Sn* attenuation (Fig. 2) (Barron &

866 Priestley, 2009). Although *Sn* is clearly strongly excited for S3, the *Sn* energy arrives towards the

867 end of the *Sn* window (Fig. 11). This likely represents a delayed *Sn* arrival rather than

868 incorporation of early *Lg* into the *Sn* window, because early *Lg* should be followed by stronger

869 subsequent *Lg* waves (Fig. 4, f&g) yet the energy in the *Sn* window is already the strongest in the

870 entire record. Because our standard *Sn* window does not capture much of the *Sn* wavetrain for

871 S3, inevitably *Sn/Lg* – calculated as the ratio of the RMS amplitudes of the respective windows –

872 is lower than expected, explaining why only 39% of stations record *Sn/Lg* > 2. This analysis, and

873 the clear increase in *Sn/Lg* for stations ~100 km north of the ramp (Supplementary materials S8)

874 persuade us that S3 is indeed a mantle earthquake. The S1 and S6 record sections are quite

875 different from S2, S3, S4 and S5. Neither S1 nor S6 shows significant *Sn* excitation relative to

876 *Lg* excitation, and they do not show increase in *Sn/Lg* for stations ~100 km north of the ramp

877 (Supplementary materials S8), further corroborating their crustal origin.

878

879 The *Lg* wavetrains for shallow events S1 and S6 have rather uniform amplitudes across the

880 HiCLIMB array, but *Lg* varies dramatically for likely below-Moho events S2–S5. A common

881 pattern for S2–S5 is that the southernmost few traces (<~15 km beyond YZS for S2 & S3, and

882 <~100 km beyond YZS for S4 & S5) have *Lg* wavetrains comparable to, or even larger than (S3

883   and S4) their respective *Sn* wavetrains; then the *Lg* wavetrain becomes uniformly low amplitude

884   further north. We believe this pattern may be a signature of enhanced *Lg* due to *Sn*-to-*Lg*

885   conversion at the ramp. If true, it means *Sn*-to-*Lg* conversions waves may not persist in the crust

886   for long distances, and may attenuate much faster than predicted by our modelling (which uses a

887   scatterer-free crust). Note that the relative change of *Lg* amplitudes across the array that is

888   obvious for events S2–S5 in their record sections, i.e. by comparison within traces (Fig. 11), is

889   not obvious when looking only at the array-normalized *Lg* amplitudes (Fig. 10b), which are

890   essentially the same as absolute amplitudes.

891

892   **5.3 *Lg* HF/LF, ratio of *Lg* amplitudes at higher and lower frequencies**

893   Another possibility to identify *Sn*-to-*Lg* conversion in real data, instead of relying on observing

894   an increase of *Lg* amplitudes that can be strongly influenced by factors such as site effects (Fig.

895   10b), is the enrichment of high-frequency (HF) *Lg*. We analyze our twelve HiCLIMB

896   earthquakes and four Gangdese-92 earthquakes exactly as we processed our synthetics. We have

897   no measurements from within the *Sn*-attenuation region (Fig. 2): Gangdese-92 did not extend

898   into this area, and the HiCLIMB stations here all lack high-quality low-frequency (LF) data. For

899   *Lg* from the six southern events (Fig. 12a), we see the southern few stations, in particular those

900   within the ramp region (negative distances), do have a much larger high-frequency component

901   compared to the more northern stations, where *Lg* HF/LF ratio is more uniform. The peaking of

902   *Lg* HF/LF may be smaller for the crustal events (open symbols) than for the mantle events

903   colored symbols, Fig. 12a), as predicted by synthetics (Fig. 8a). For the four events recorded on

904   the Gangdese-92 array, we more clearly see the rise of *Lg* HF/LF associated with the end of the

905   ramp (Fig. 12c) because there are more stations vertically above the Moho ramp. However, we

906   do not see an *Lg* HF/LF peak associated with the end of the Moho ramp for the six northern

907   events (Fig. 12b), because these events have not traversed the ramp.

908   # 6. Discussion

909   We now bring together our numerical and observational results, to address our three main

910   results: the ability to use *Sn/Lg* to recognize below-Moho earthquakes even in the presence of

911   significant crustal thickening, our identification of *Sn*-to-*Lg* conversion in real data, and the

912   value of *Lg* frequency content as another discriminant for continental mantle earthquakes.

913

914   Our numerical results (Figs. 6–9) show that significant Moho topography, that locally enhances

915   *Sn* amplitudes and more regionally enhances *Lg* amplitudes, does not strongly influence *Sn/Lg*

916   ratios which remain useful as a comparative measure to separate mantle and crustal earthquakes.

917   The resilience of the *Sn/Lg* method to crustal thickening is clear because *Sn/Lg* ratios for the

918   deeper-lid (65-km) and shallow-lid (35-km) events are always above the *Sn/Lg* ratios for the

919   mid-crustal earthquake (15-km) at the same distance (Figs. 6g,h,i, 9g,h,i). The best separation, an

920   order of magnitude, occurs between our shallow-lid earthquake and our mid-crustal earthquake

921   at stations far beyond the end of the ramp, because of the ramp-transient nature of *Sn* amplitude

922   perturbations and modest *Sn*-to-*Lg* conversion for shallow-lid earthquakes.

923

924   Thus our simulation results show we can apply *Sn/Lg* criteria to identify mantle earthquakes

925   regardless of the presence of a Moho-thickening ramp. Observations of *Sn* and *Lg* on the

926   HiCLIMB (Fig. 10 a&b) and Gangdese-92 arrays (Supplementary S7) show less significant

927   effects than our simulation results (Figs. 6&9, right columns) that therefore likely represent the

928   strongest possible scenarios for ramp effects on *Sn/Lg* signatures. Our HiCLIMB events are

929   strongly influenced by site effects and are obliquely incident on the array (though as noted

930   above, this obliquity is likely unimportant), whereas our Gangdese-92 events do not exhibit

931   strong site effects and are nearly in-line with the array. Nonetheless, neither set of events shows

932   either the predicted strong focusing of *Sn* near the end of the ramp nor the predicted sustained

933   increase of *Lg* energy beyond the end of the ramp (Fig.10, Supplementary materials S7). We

934   believe these inconsistencies between data and simulations originate from the absence in our

935   models of small-scale features such as inhomogeneities in the crust or less-smooth Moho

936   topography. Additional small-scale features should spatially smooth a localized feature such as

937   the *Sn* peak (Figs. 6&9, left column), and selectively attenuate the higher-frequency *Lg* in the

938   crust produced by *Sn*-to-*Lg* conversion (Fig. 8), which we discuss more below. Our observations

939   on individual *Sn* and *Lg* waves agree with findings in the North Sea (Mendi et al., 1997) that

940   regional waves are more influenced by small-scale scatterers than large-scale features. Because

941   the largest perturbations from the reference model due to a Moho ramp are the *Sn* peak above

942   and the increased *Lg* beyond the ramp, smoothing out these effects in the real data likely means

943   *Sn/Lg* in the real world is even more robust than predicted by our simulations.

944

945   We can directly compare *Sn/Lg* for events traversing one of Earth's largest Moho ramps with

946   *Sn/Lg* for events traversing relatively uniform Moho topography (Figs. 2, 9e). Using a previously

947   established *Sn/Lg* threshold that identified two new below-Moho earthquakes in NW Tibet

948   (Wang and Klemperer, 2021), we can identify four earthquakes (S2, S3, S4, S5) south of the

949   MCT that nucleated below the Moho, including one previously tentatively identified as such

950   (S3=H82 of Baur, 2007) and one that has a nominal (assigned) catalog depth of 10 km (S5). We

951   can similarly show that a different event with a catalog depth close to the Moho (S1) is in fact a

952   crustal event. We emphasize that these conclusions are quite reliable, as they are based on

953   measurements on multiple stations that show *Sn/Lg* significantly larger than the regional low-

954   *Sn/Lg* baseline established for multiple nominally shallow earthquakes in both northern India and

955   in northwestern Tibet.

956

957   *Sn*-to-*Lg* converted waves maybe most easily identified in the frequency domain (Fig. 12), rather

958   than in the amplitude domain (Figs. 10&11), through *Lg* HF/LF. This diagnostic is motivated by

959   early mode-coupling studies (Maupin, 1989) (Fig. 3) and verified with our full-waveform

960   synthetics (Fig. 8). Two groups of events with significant Moho ramp crossing recorded on two

961   separate arrays both exhibit increase of *Lg* HF/LF (Fig. 12a&c) associated with the end of the

962   ramp, but another group of non-Moho-ramp crossing events recorded on one of the same arrays

963   does not show this (Fig. 12b). Hence, we believe *Lg* HF/LF is a rather robust signature of *Sn*-

964   converted-*Lg* waves. This implies that the enhanced *Lg* above and close to the ramp on the

965   record sections of the southern mantle events S2–S5 (Fig. 11) represents *Sn*-to-*Lg* conversions,

966   enriched in high-frequencies. The enriched HF content for *Lg* close to the ramp corroborates our

967   suspicion that small-scale crustal scatterers are the reason we do not see persistent high *Lg*

968   energy after conversion in real data, unlike in the numerical results.

969

970   A prominent feature of *Lg* HF/LF is the clear separation of mantle and crustal earthquakes

971   recorded on HiCLIMB (Fig. 12a&b) following our interpretations based on *Sn/Lg* (Fig. 10e),

972   whereas the overlapping of *Lg* HF/LF for the Gangdese events (Fig. 12c) matches their

973   overlapping *Sn/Lg* values (Supplementary materials S7). This can be understood from a normal-
974   mode perspective in that the only *Lg* energy excitable by a mantle earthquake is associated with
975   lower-frequency Airy phases (Knopoff et al., 1973) that could have a displacement/strain
976   eigenfunction sampling the mantle lid to some depths (Wang and Klemperer, 2023, their Fig.
977   3a), whereas the higher-frequency *Lg* Airy phases have displacement/strain eigenfunctions much
978   more tightly bounded within the crust (Wang and Klemperer, 2023, their Fig. 3b). In our
979   reference model with flat Moho at 30 km, a source 5 km above Moho (z=25 km) has essentially
980   the same *Lg* HF/LF as a source 15 km above Moho (z= 15 km), and both are clearly distinct
981   from *Lg* HF/LF for a source that is 5 km below the Moho (z=35 km). The deeper-lid earthquakes
982   at $z = 65$ and 95 km have similar *Lg* HF/LF as the crustal earthquakes at short offsets because of
983   the artificial inclusion of *Sn* in our measurement windows (Fig. 4 a,c,d,e), but their *Lg* HF/LF
984   quickly drops beyond ~400-500 km epicentral distance as *Sn* exits the *Lg* window. Hence, like
985   *Sn/Lg*, *Lg* HF/LF is not particularly sensitive to absolute source depths, but rather to their relative
986   position with respect to the Moho as predicted by the normal-mode explanation, so can also be
987   used as a discriminant for sub-Moho earthquakes. This frequency discriminant *Lg* HF/LF is even
988   simpler than *Sn/Lg* because *Lg* HF/LF remains almost constant with epicentral distance (Fig.
989   13a) (apart from the artificial sinusoidal oscillations due to overlapping *Lg* and *Sn* windows, see
990   Section 3) in contrast to *Sn/Lg* (Fig. 5), and because *Lg* in general is a much simpler wave than
991   *Sn* (i.e. crustal waveguide for *Lg* vs. whispering-gallery waveguide for *Sn*).

992

993   We further selected crustal thickening models with small *d*'s and *w*'s in order to capture the
994   effects a Moho thickening ramp can produce on *Lg* HF/LF (Fig. 13 b-e). The difference between
995   *Lg* HF/LF for a crustal and an upper-mantle earthquake is present for all our selected models
996   (beyond ~400–500 km where some *Sn* is present in the *Lg* window), representing among the
997   strongest effects a Moho ramp can produce. We note *d* has a stronger effect than *w* in terms of
998   increasing the upper-mantle event's high-frequency *Lg* thereby raising its *Lg* HF/LF, and when
999    $d = 100$ km, the separation with the mid-crustal event is quite small (Fig. 13 b,d&e). In
1000   addition, our synthetics show overlapping *Lg* HF/LF for the deeper-lid event and the mid-crustal
1001   earthquake in these extreme models (Fig. 13 b–e). However, these are a worst-case because the
1002   high-frequency enriched *Lg* due to *Sn*-to-*Lg* conversion, that is persistent at large distances in our
1003   synthetics (Fig. 8), in real data fades away quickly after the end of the ramp as observed in data

1004 (Figs. 11&12). Hence, it is unlikely *Lg* HF/LF will be undistinguishable for upper-mantle and

1005 mid-crustal earthquakes nor will it mis-classify a deeper-lid earthquake as a crustal earthquake if

1006 measurements are made on sufficient stations beyond the end of the ramp. In ongoing work, we

1007 are exploring the correlation between *Lg* HF/LF and *Sn/Lg* amplitude-ratio discriminants, and

1008 their joint potential to resolve relative location of earthquakes above and below the Moho.

1009

7. Conclusion

1011 We enhanced the code AxiSEM3D to perform 2.5D regional wave simulations across a Moho

1012 ramp and achieved a combination of higher frequency ranges and longer propagation distances

1013 than other recent studies. Most notably, our modifications enabled checking the representation of

1014 an undulated geometry within AxiSEM3D and using this technique to stretch a uniform mesh so

1015 that the computed wavefield can be shown at the correct positions, avoiding wavefield

1016 distortions that will be visible for simulations at our scale (i.e. regional, vs. global).

1017

1018 We compare our numerical results in a 1D reference model, with flat Moho, with previous

1019 studies on *Sn* and *Lg* geometrical spreading to confirm the accuracy of our numerical approach.

1020 In addition, with this benchmarking exercise we emphasize the fact that regional-wave arrival

1021 windows, as defined by group velocities, cannot be fine-tuned in real data. The windows will

1022 always overlap leading to artificial abrupt or oscillatory changes in measured amplitudes and

1023 frequencies whenever an *Sn* or *Lg* phase moves in or out of its window (a phenomenon

1024 previously noted by Yang (2002)). It is likely that mischaracterization of phases contributed to

1025 an over-estimation of *Sn* amplitude increase at ~700–1,300 km by Yang et al. (2007) leading to

1026 an inaccuracy in their *Sn* geometrical spreading model (Fig. 4b).

1027

1028 As we vary distance to ramp start d and ramp width w in our crustal thickening model (Fig. 1),

1029 the synthetics for *Lg* absolute amplitudes are relatively simple and consistently display sustained

1030 increase *Lg* for amplitudes as well as *Lg* HF/LF across the ramp, though to the smallest degree

1031 for the crustal source. On the other hand, synthetic *Sn* absolute amplitudes are much more

1032 complicated due to its complex propagation path as an interference head wave. Nonetheless,

1033 commonalities are present, including the *Sn* focusing peak around the ramp end, and the return

1034   to *Sn* amplitudes similar to the reference model at larger distances for almost all parameter

1035   ranges tested. These phenomena are closely related to the shape of the *Sn* waveguide (Fig. 7).

1036   Even with the presence of these perturbations on individual amplitudes, among all cases tested

1037   in our simulations, *Sn/Lg* ratios for mid-crustal earthquakes are persistently lower than for

1038   mantle earthquakes on noise-free synthetics, and potential confusions are unlikely when using a

1039   recording array with varying source-station geometry.

1040

1041   There are substantial differences between real-world data and synthetics for individual *Sn* and *Lg*

1042   absolute (or array-normalized) amplitudes, as in addition to factors like site effects, the real

1043   world contains many finer-scale details, such as crustal scatters and irregular Moho/ramp

1044   surfaces that tend to average the sharp *Sn* focusing peak and sustained *Lg* amplitude increase

1045   seen in our synthetics. We therefore believe our ramp models provide a worst-case scenario for

1046   the utility of *Sn/Lg* in the real world as waveforms are smoother in the real world. We verified

1047   the effectiveness of *Sn/Lg* through direct comparison with ramp-crossing and non-ramp-crossing

1048   events from southern and northwestern Tibet, recorded on the same array with roughly the same

1049   epicentral distances (Fig. 10e; Supplementary materials S6), providing strong evidence for four

1050   mantle earthquakes in northern India.

1051

1052   *Sn*-to-*Lg* converted waves are generally hard to recognize from their amplitudes, though not

1053   impossible (Fig. 11), but can more easily be identified by the shift of *Lg* frequency content, as

1054   shown here with our full-waveform synthetics (Fig. 8) and demonstrated with non-ramp-crossing

1055   events (Fig. 12b) and two sets of ramp-crossing events recorded on two different arrays (Fig.

1056   12a,c). *Lg* HF/LF is a promising new discriminant to identify continental mantle earthquakes

1057   from their decreased *Lg* HF/LF as predicted by normal-mode theory and verified in both our

1058   reference and ramp models (Fig. 13).

1059

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CREDIT statement

Shiqi Wang:

Conceptualization, Methodology, Software, Validation, Formal analysis, Investigation, Data curation, Writing-Original draft, Writing-Review and editing, Visualization.

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Methodology, Validation, Formal analysis, Investigation, Writing-Review and editing, Visualization, Funding acquisition.

1079

Data Availability Statement

All seismic data analyzed in this paper are available via

https://www.fdsn.org/networks/detail/XF_2002 (HiCLIMB data) and at

https://doi.org/10.5281/zenodo.10971752

(Gangdese-92 data). Our custom-version AxiSEM3D can be found at

https://github.com/axelwang/AxiSEM3D_Modified.

1086

Competing interest

The authors declare no competing interest.

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1400 **Figure and table legends**

1401

1402 **Table 1. Earthquakes recorded on the HiCLIMB array.** Southern events are named S1–S6.
1403 These events nucleated in northern India and cross a significant Moho ramp before being
1404 recorded by HiCLIMB (Fig. 2). The distance to the ramp (*d*) and ramp-width (*w*) are shown as
1405 ranges because of the different azimuth (hence obliquity to the ramp) from each earthquake to
1406 the southern and northern limits of the HiCLIMB stations (Fig. 2). The six events with no values
1407 for *d* and *w* comprise our 'northern' events that do not cross significant Moho topography before
1408 reaching the stations (Wang and Klemperer, 2021). Magnitude and depth data are from PDE,
1409 2024. Values in parentheses from Baur (2007). Italicized hypocentral locations and depths are
1410 from the Seismological Bulletins of the Indian National Center for Seismology.

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1413 **Fig. 1. Computational model and a representative wavefield.** Computational region extends
1414 to 2000 km in range and 230 km in depth. Thick black line shows the Moho, which is at 30 km
1415 on the left side, and transitions smoothly to 60 km through a 30-km high ramp whose width (*w*)
1416 and distance from source (*d*) are labelled. Small red stars represent the 3 source depths (*z*) we
1417 study, 15, 35 and 65 km, respectively. Thin black lines represent the top and bottom of the
1418 mantle low-velocity zone (LVZ) at 80 and 220 km. In this example the source is at 65 km depth.
1419 A snapshot wavefield (transverse component, filtered 1-5 Hz) is plotted at time 235.5s with
1420 amplitude shown in the color bar on lower left, showing multiply-reflected and interfering
1421 regional wave trains. The wavefield in the crust is complex as it is a combination of multiple
1422 reflections from the Moho top-side (*Lg*), under-side (*Sn*), as well as from just below the LVZ
1423 (*Sa*). The absence of visible reflections from the bottom of the computational domain, despite the
1424 clearly visible reflections from the bottom of the LVZ, demonstrates the performance of our
1425 absorbing boundary condition.

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1428 **Fig. 2. Earthquakes and stations.** (a) Earthquakes in India south of the Main Frontal Thrust
1429 (MFT) and in the Bhutan Himalaya (red stars, red labels S1–S6; Table 1) recorded on the
1430 HiCLIMB array (Nabelek et al., 2009) (purple triangles) after crossing a ~15–20 km high Moho
1431 ramp from thinner to thicker crust. Earthquakes in northwestern Tibet (black stars) (Wang and
1432 Klemperer, 2021) are recorded on the same array but their paths do not cross significant Moho
1433 undulations. Moho depths are interpolated from CRUST1.0 to 0.05° (Laske et al., 2013).
1434 Gangdese-92 array (Shi et al., 2015) (yellow triangles) recorded four nominally deep earthquakes
1435 ~ due south of the array (yellow stars, yellow labels G1-G4) (Supplementary materials table S7-
1436 1). YZS: Yarlung-Zangpo, BNS: Banggong-Nujiang, JRS: Jinsha River sutures. H: Himalaya, L:
1437 Lhasa, Q: Qiangtang, SG: Songpan-Ganzi terranes. MFT: Main Frontal thrust, KKF: Karakoram
1438 fault, KXF: Karakax fault. White dashed lines border a well-known attenuation zone for *Sn* (e.g.
1439 Barron and Priestley, 2009). Black double arrow indicates approximate distance from YZS to the
1440 attenuation zone. YZS represents the ending of the Moho ramp, while two thick green lines
1441 represent the Moho ramp beginning directly south of the arrays at the approximate location for

1442   Main Central Thrust (MCT), which is too tortuous to show on our map (see main text). (b).
1443   Cartoon crustal and Moho cross-section along white solid line shown in (a), redrawn after
1444   Nabelek et al., 2009 based on their receiver function analysis on the HiCLIMB array.
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1447   **Fig. 3. Normal-mode-coupling results.** Transmission amplitude coefficients for Rayleigh waves
1448   due to perpendicular incidence onto a North Sea graben-type model, visualized from Figure 4 of
1449   Maupin (1989). The matrix is symmetric; for convenience we label each row as representing the
1450   incidence of a pure mode and each column as the converted amplitude with the amplitude
1451   coefficients representing the degree of partitioning of energy due to incidence onto a large-scale
1452   Moho depth variation. The calculations are done for 1 Hz, at which a strict separation (shown as
1453   black dashed lines) can be made between *Lg* (mode numbers ≤ 11) and *Sn* (mode numbers ≥ 12).
1454   Note the amplitude coefficients are typically large along the diagonal (no mode conversion), and
1455   are very small in the upper right and lower left sections of the figure as separated by the dashed
1456   lines. Because *Lg*-to-*Sn* coupling is strongest into the lowest *Sn* modes (12, 13, 14) *Lg*-to-*Sn*
1457   coupling preferentially excites the lower frequencies of *Sn*. For example, looking at the row for
1458   mode 9, the squares in columns 1-11 represent mode coupling to other *Lg*-forming normal
1459   modes, though most of the energy remains as mode 9 (highlighted with thick black border).
1460   Across the dashed line, squares in columns 12-25 represent mode coupling into *Sn*-forming
1461   normal modes, leading to *Lg*-to-*Sn* conversion with the strongest coupling to mode 13
1462   (highlighted with dashed border), a low mode number for *Sn* normal modes. Similarly, *Sn*-to-*Lg*
1463   coupling is dominantly from the lower *Sn* modes (e.g. 12, 13, 14) to the higher *Lg*-forming-
1464   modes (e.g. 8, 9, 10, highlighted with dotted lines, contributing dominantly to higher-frequency
1465   *Lg* Airy phases.
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1468   **Fig. 4. *Sn* and *Lg* in reference model (no ramp).** Transverse-component displacements are
1469   shown. (a) *Lg* amplitude filtered 1–5 Hz, for three source depths, "mid-crustal" (15 km),
1470   "shallow-lid" (35 km) and "deeper-lid" (65 km). (b) *Sn* amplitude at 3 Hz for the same three
1471   source depths. Black lines in (a) and (b) are the best-fit models of Yang (2002) (*Lg*) and Yang et
1472   al. (2007) (*Sn*) that only predict relative amplitudes as a function of distance, so are set to be
1473   equal to our results at 200-km distance for *Lg*, and 300 km for *Sn* (the starting modelling distance
1474   in Yang et al. (2007). Our extrapolation of the Yang (2002) and Yang et al. (2007) formulae
1475   beyond the distance range they studied leads to large misfits at large offsets. Both (a) and (b) are
1476   log-log, amplitude vs. distance. Seismograms for symbols with black border are shown in (c)-(i)
1477   with corresponding labels. Two red vertical lines bound the *Sn* windows, and cyan lines bound
1478   the *Lg* windows. Full *Lg* windows are not shown for (f)-(i) as the focus there is on the *Sn*
1479   window. Seismogram amplitudes shown are absolute values without normalizations, in units of
1480   nanometers (nm).
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1483   **Fig. 5. *Sn/Lg* in reference model (no ramp).** *Sn/Lg* for three source depths, 15-km (mid-crust),
1484   35-km (shallow-lid) and 65-km (deeper-lid) are clearly separated at all epicentral distances,

1485   despite their individual variations with offset. A black fiducial line at *Sn/Lg* =0.2 further
1486   illustrates separation of crustal and mantle earthquakes.
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1488   **Fig. 6. *Sn*, *Lg* amplitudes and *Sn/Lg* with varying distance to ramp start *d* but fixed ramp**
1489   **width *w* = 200 km.** Rows, top to bottom, display results when the source is mid-crustal (15 km),
1490   shallow-lid (35 km) and deeper-lid (65 km). Columns, left to right, show amplitudes of *Sn* and
1491   *Lg* relative to the reference model, and *Sn/Lg*. *Left and middle columns* are plotted with data
1492   aligned at the ramp, with its beginning marked as a vertical dashed black line and end marked as
1493   a vertical solid black line (0 on the horizontal axis). A grey horizontal line at 1 marks no
1494   deviation from reference-model results. The vertical axis is plotted in $\log_{10}$ scale while the
1495   horizontal axis is linear. Note because of the ramp alignment and a fixed total simulation range,
1496   larger *d* has a shorter distance covered beyond the end of the ramp. *Right column* plots *Sn/Lg*
1497   against epicentral distance and superimposed on the reference-model results (grey inverted
1498   triangles). Each colored bar represents the end of the ramp for the correspondingly colored
1499   symbol (e.g. the red bar marks the end of the ramp at 300 km epicentral distance for *d* = 100 km
1500   (red circles), and its ramp starts outside the range of the plots). The total ramp ranges for the
1501   other cases are shown between the vertical lines (e.g. for *d* = 300 km (blue diamonds) the ramp
1502   range is between the red and blue bars). Both the vertical and horizontal axes are in $\log_{10}$ scale. A
1503   solid black line at *Sn/Lg* = 0.2 (as in Fig. 5) in all rows in the right column shows that despite the
1504   variability within each plot, *Sn/Lg* for our mantle earthquakes (h) and (i) is greater than *Sn/Lg* for
1505   our crustal earthquake (g) at every common offset.
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1509   **Fig. 7. Disruption of *Sn* waveguide by the Moho ramp.** (a)-(d) Representative wavefield
1510   snapshots (for *z* = 35 km, corresponding to the blue diamonds in Fig. 6 b,e,h) bandpass filtered
1511   1-5 Hz and shown with same color scale for the amplitudes. (a) and (c): *d* = 300 km, *w* = 200
1512   km; (b) and (d): reference model (flat Moho). UT: displacement on the transverse component, in
1513   meters. Moho and top of LVZ are marked by thick and thin black lines. At time = 73 s, before
1514   the wavefield interacts with the ramp, the ramp model (a) and reference model (b) show the same
1515   wavefield, with the first arrival due to a rather weak transmitted wave through the Moho from
1516   the leading strong sub-Moho wavefront, showing an effective *Sn* waveguide. At time = 100 s,
1517   after the wavefield in the ramp model starts to interact with the ramp, the transmitted wave
1518   becomes much stronger in the ramp model (c) than in the reference model (d) (red arrows)
1519   corresponding to the onset of the *Sn* peak right after the vertical black line (0 km, end of the
1520   ramp) in Fig. 6b. (e) Schematics of sub-Moho wavefront interacting with a flat Moho (black
1521   line/blue raypath) and with a Moho with a thickening ramp (grey line/dashed yellow raypath). (f)
1522   calculated energy partitioning for a transverse S-wave incident on the Moho from below. In (e),
1523   black dot on the red wavefront represents a point slightly below the flat Moho that will
1524   contribute to *Sn* for the reference model where the purple arrow intercepts the Moho with
1525   incidence angle, i_{flat} close to 90°, suggesting most of the energy is reflected back below the
1526   Moho (dashed black curve in f), representing the *Sn* waveguide, while a smaller amount is
1527   transmitted into the crust at a smaller angle r_{flat}. The introduction of a Moho ramp reduces these
1528   angles to i_{ramp} and r_{ramp}, as shown by the yellow dashed arrows, and sharply increases the
1529   amount of energy transmitted into the crust (black curve in f). Because r_{ramp} is smaller than

1530 r_{flat}, the horizontal velocity (apparent velocity) of *Sn* is reduced in the ramp region. Points on
1531 the wavefront deeper than the black dot will not interact with the ramp, but will enter the
1532 thickened crust beyond the end of the ramp, thus explaining the recovery of *Sn* amplitude at
1533 distances further away from the end of the ramp. (g) seismogram at 510 km (10 km beyond the
1534 ramp and approximately corresponding to the largest amplitude peak in Fig. 6b) for the reference
1535 model (black line) and ramp model (blue line) with *Sn* and *Lg* windows marked by red and cyan
1536 lines, showing the large growth of the first-arrival *Sn* wave and the phase delays experienced by
1537 the ramp model.
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1540 **Fig. 8. Change of *Lg* frequency content as a result of *Sn*-to-*Lg* conversion.** Same as in the
1541 middle column of Fig. 6, but the vertical axis '*Lg* HF/LF' is the ratio of high-frequency (HF, 1–5
1542 Hz) *Lg* to low-frequency (LF, 0.1–0.8 Hz) *Lg* of the ramp model divided by the equivalent ratio
1543 for the reference model. For all panels, the horizontal axis is linear while the vertical axis is in
1544 log scale.
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1547 **Fig. 9. *Sn*, *Lg* amplitudes and *Sn/Lg* with varying ramp width *w* but fixed ramp distance *d***
1548 **= 300 km.** Figure organization as for Fig. 6, except for the left and middle columns the vertical
1549 colored bars represent the start of the ramps for the correspondingly colored symbols. The end of
1550 the ramp is aligned for all of these cases at 0 km and marked by a solid black line, as in Fig. 6.
1551 For the right column, the beginning of the ramp is marked by a dashed black line and the end of
1552 the ramp is marked by a solid colored line for the corresponding colored symbol. For example,
1553 the ramp region for the red circles is within the dashed black line and the solid red line. For all
1554 panels, the vertical axis is in log scale. Horizontal axes are linear for parts (a)–(f), and log scale
1555 for (g), (h), (i).
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1558 **Fig. 10. *Sn*, *Lg* amplitudes and *Sn/Lg* as a function of south-north distance from ramp end.**
1559 Southern events are shown with colored symbols and northern events with black symbols. Events
1560 interpreted as mantle earthquakes are shown with solid symbols and crustal events with open
1561 symbols. The beginning of the ramp, perpendicular to the array, is shown with a dashed black
1562 line and the end of the ramp is shown with a solid black line. The southern limit of the *Sn*
1563 attenuation zone is marked with a magenta dashed line. (a)-(d) Individual *Sn* and *Lg* amplitudes
1564 for the southern and northern events, respectively. Data points are aligned vertically for each
1565 individual station location. (e) Comparison of *Sn/Lg* for the northern and southern events. This
1566 very different combined group of events can be clearly separated by high and low *Sn/Lg*,
1567 especially >~100 km beyond the end of the ramp (Supplementary materials S7). For all panels,
1568 the horizontal axis is linear while the vertical axis is in log scale.
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1572 **Fig. 11. Record sections of the southern events.** Top four events (S2–S5) are interpreted here
1573 as below-Moho earthquakes, and bottom two events (S1, S6) as crustal events. For each event,
1574 an upper panel shows *Sn/Lg* measured at each station, on a linear scale from 0-8, with our

1575 arbitrary threshold $Sn/Lg = 2$ shown as a grey line. The event code along with the percentage of
1576 stations that registered a $Sn/Lg > 2$ are labelled in the upper panel. For each event, between the
1577 two panels is each station's distance north of the YZS, recognized as the end of the Moho ramp.
1578 The bottom panels show trace-normalized amplitudes for each event, with Sn window colored
1579 red, Lg window cyan, and the noise window green. The traces are shown with a reduction
1580 velocity of 4 km/s. Traces are displayed south to north with epicentral distance shown beneath
1581 each record section. Thick yellow line marks stations south of YZS (i.e. within the ramp), and
1582 thick magenta line marks stations within the Sn attenuation zone.

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1585 **Fig. 12. High-frequency (HF) to low-frequency (LF) ratio of *Lg* waves in data.** Moho ramp
1586 beginning is marked by a black dashed line while its ending is marked by a black solid line. The
1587 start of the Sn attenuation zone is shown by a magenta dashed line. The horizontal axis are
1588 distances aligned at the end of the Moho ramp. The vertical axis shows the Lg HF/LF in a log
1589 scale. (a) southern events. (b) northern events. Symbol styles are as in Fig. 10 (open symbols:
1590 crustal events; closed symbols: mantle earthquakes). (c) events recorded on the Gangdese-92
1591 array (not categorized as crustal or mantle because we lack comparison events). Note different
1592 vertical scale compared with (a)&(b).

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1595 **Fig. 13. High-frequency (HF) to low-frequency (LF) ratio of *Lg* waves in reference and**
1596 **select ramp models.** (a) Lg HF/LF for five source depths with the reference model with a flat
1597 Moho at 30 km. Two of the sources are located within the crust, one in the shallow-lid, and two
1598 deeper within the mantle. (b) & (c) Lg HF/LF for ramp models with fixed w=100 km, testing the
1599 effect of increasing d. (d)&(e) Lg HF/LF for ramp models with fixed d=100 km, testing the
1600 effect of increasing w. All panels are log-log. Dashed and solid black lines indicate the start and
1601 end of the ramp, when located beyond 200 km. A fiducial line at Lg HF/LF = 0.7 is drawn to
1602 emphasize the separation of crustal and mantle events for all panels.

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1620   **Tables**

Code name	Date, time	Location (N°, E°)	Magnitude (m_b)	Catalog depth (km)	Distance to ramp, d (km)	Effective ramp width, w (km)
S1	2005-07-26, 18:27:05	23.27, 91.41 *23.281, 91.516*	4.0	38.1± 27.4 *10*	560-660	200-280
S2	2005-05-03, 00:38:57	25.76, 91.06 *26.078, 91.033*	4.3	33.6±? *33*	285-385	210-280
S3 (H82)	2004-08-04, 02:09:21	25.92, 90.26 *25.865, 90.333*	4.2 (4.1, M_w)	61.7± 10.8 *20* (53±?)	250-330	180-260
S4	2005-05-27, 22:12:20	26.14, 87.21 *26.170, 87.685*	3.5	57.7± 12 *15*	270-280	130-160
S5	2004-11-24, 22:35:42	27.33, 90.94 *27.337, 90.875*	4.0	10±? *10*	100-140	160-380
S6	2004-08-09, 08:18:18	27.58, 91.80 *27.547, 91.718*	4.1	16.1±? *14.9*	50-120	195-475
WT1	2005-05-19, 05:43:30	35.63, 78.38	4.2	97.6 ± 14.1	-	-
WT2	2005-06-20, 22:52:26	36.23, 77.92	3.9	77.9 ± 8.4	-	-
WT3	2005-03-03, 15:07:39	35.65, 77.85	3.7	57.5 ± 16.8	-	-
04-251	2004-09-07, 04:01:05	35.72, 78.25	4.2	7.6 ± 26.8	-	-
04-291	2004-10-17, 15:35:45	35.20, 77.67	4.3	15±?	-	-
05-201	2005-07-20, 10:54:49	35.34, 77.79	4.2	10±?	-	-

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1622   Table 1

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1633   **Figures**
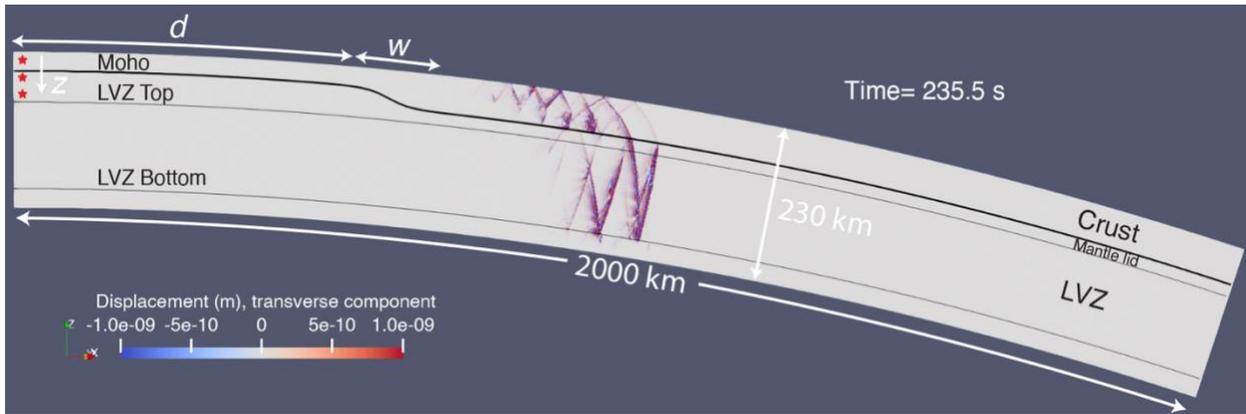
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1636   Fig. 1

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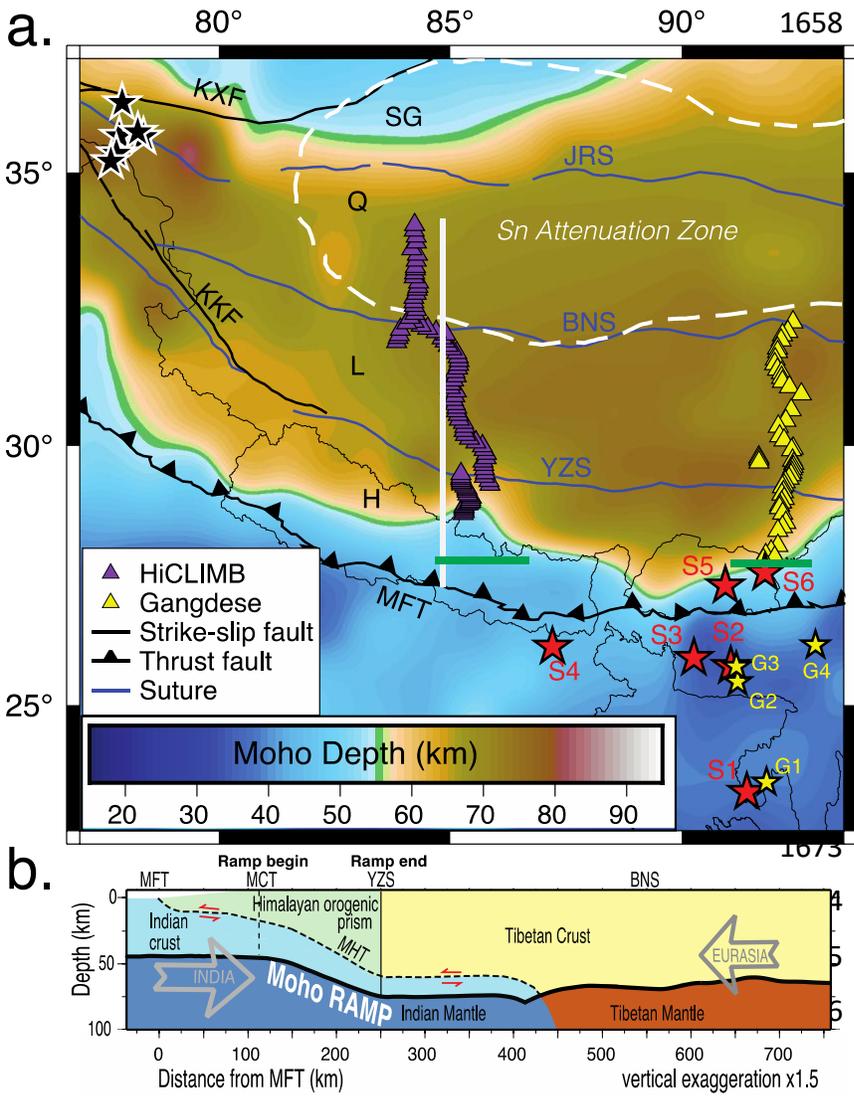
1678   Fig. 2

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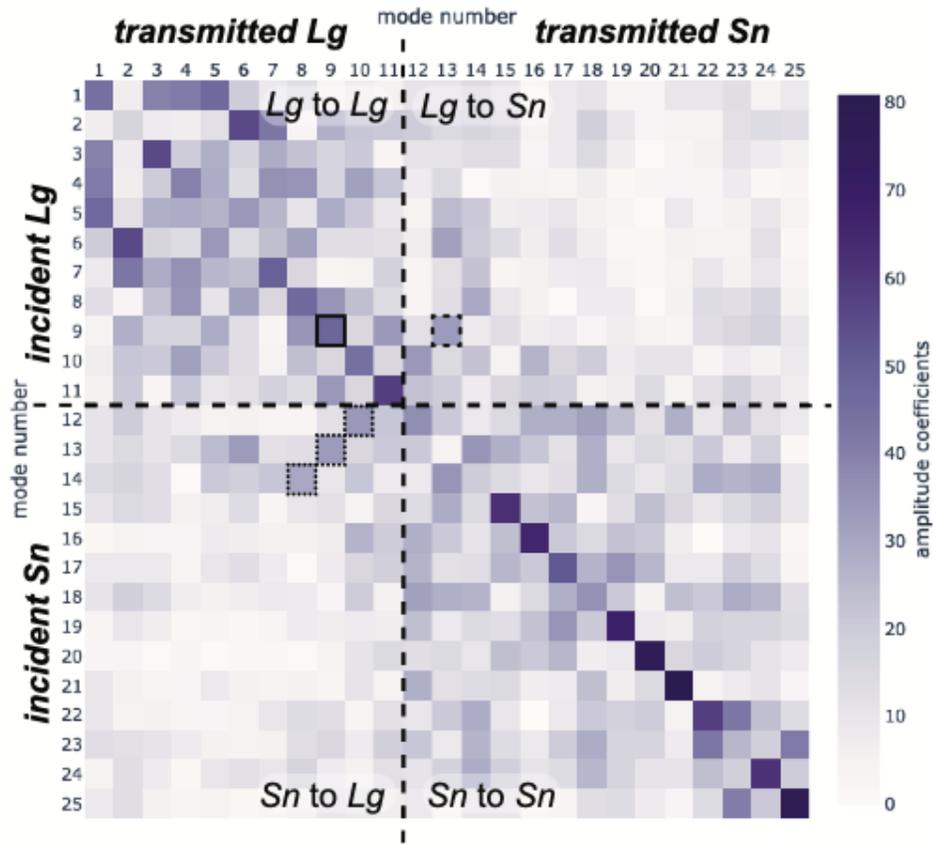
1705   Fig. 3

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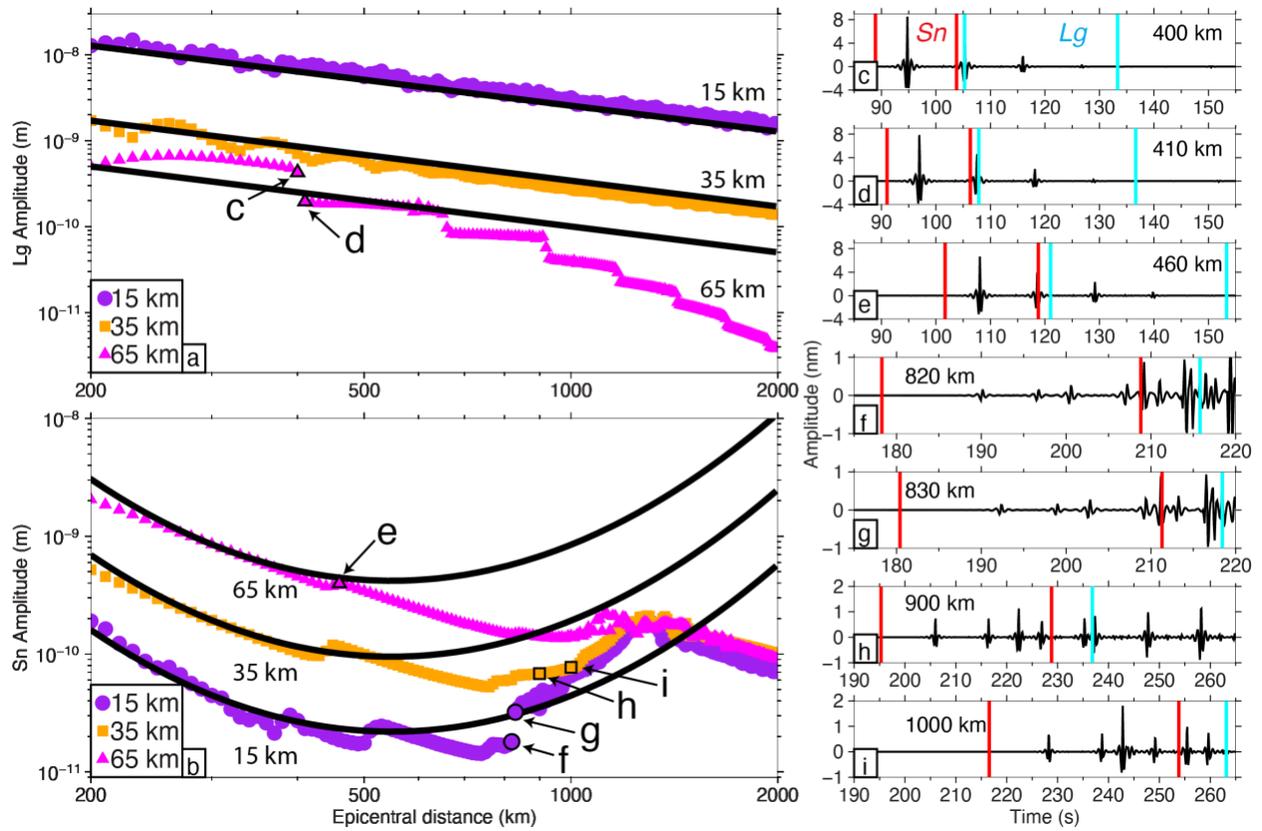
1720   Fig. 4

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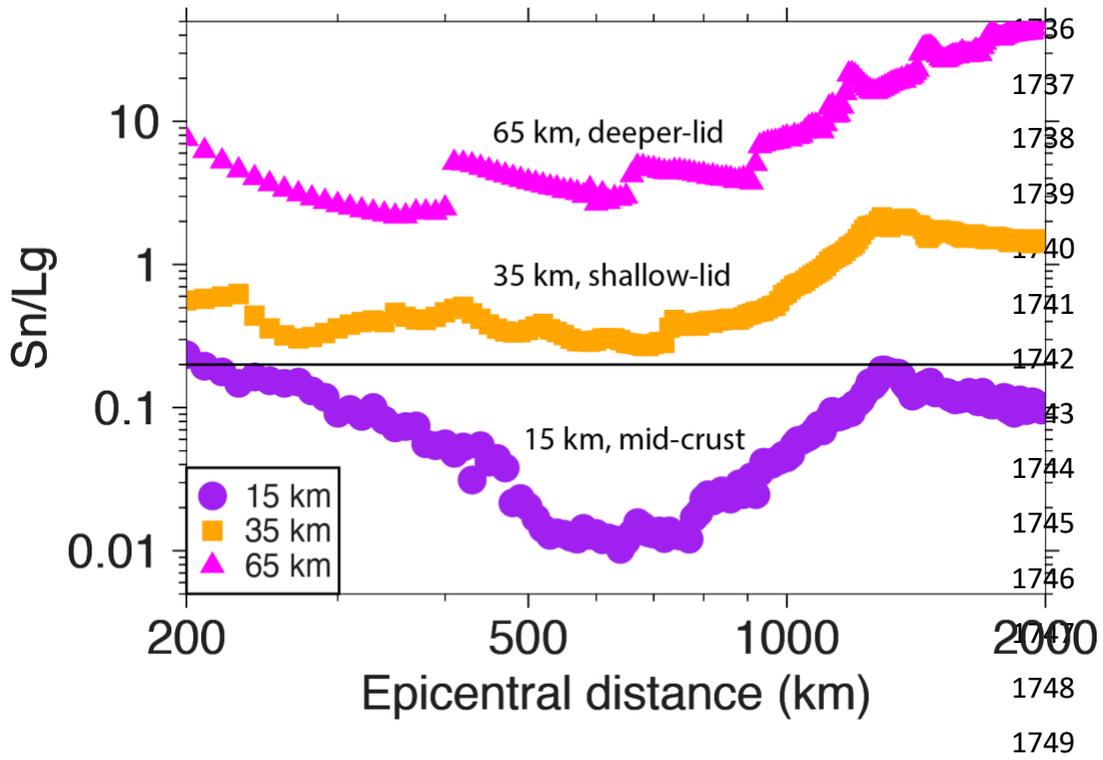
1750   Fig. 5

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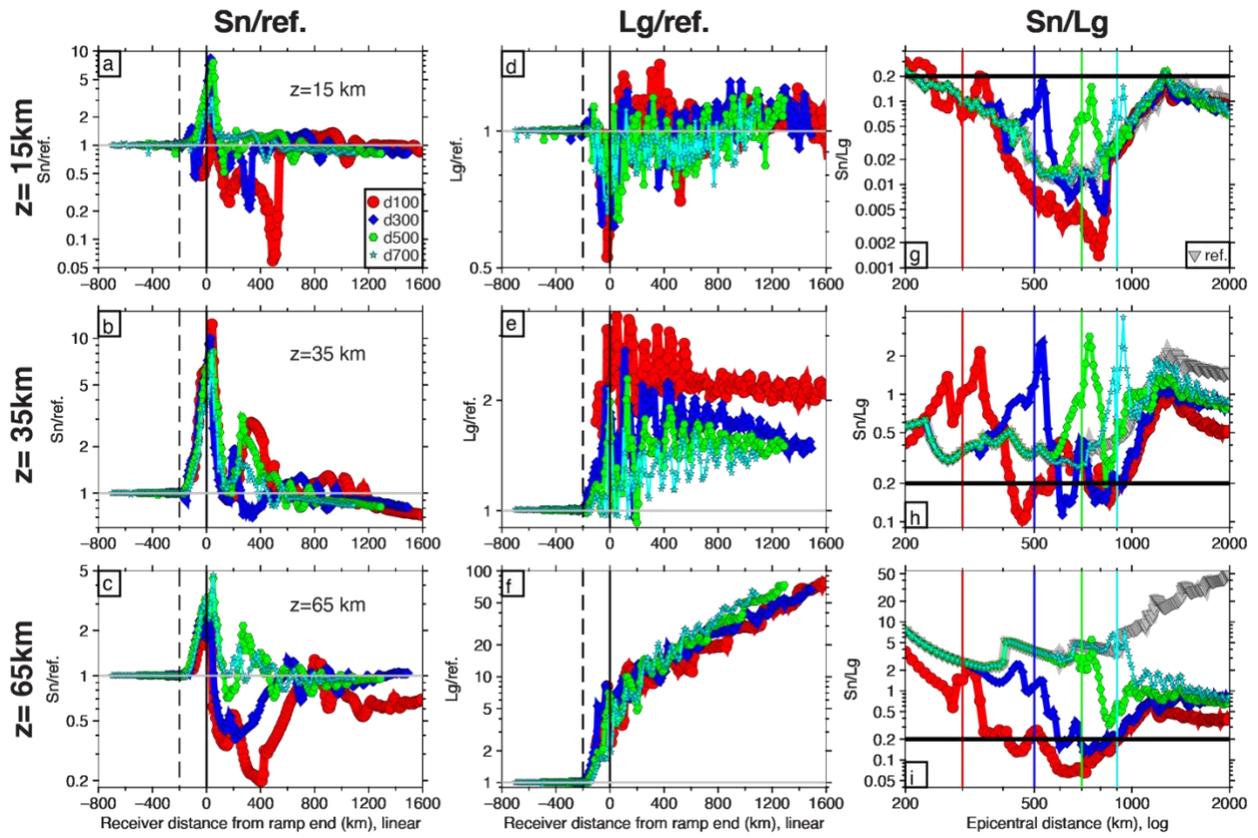
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Fig. 6
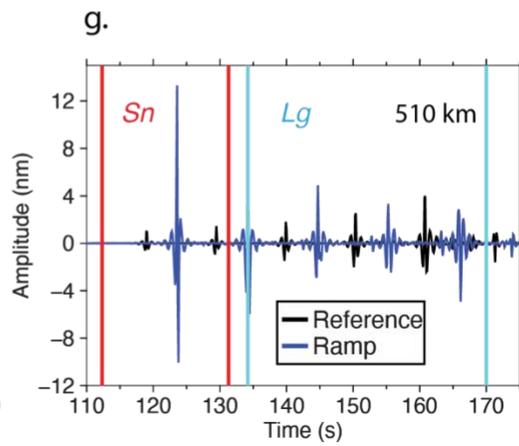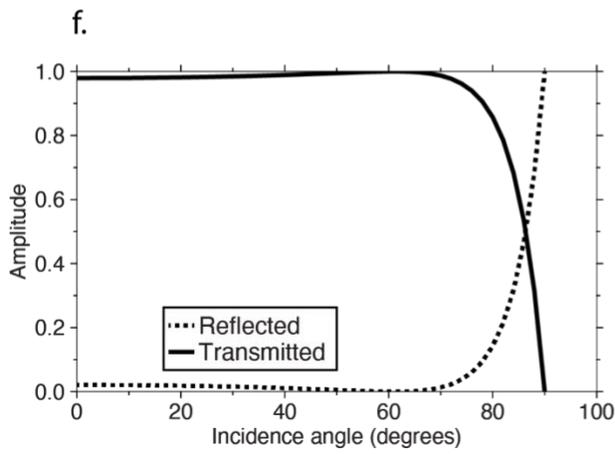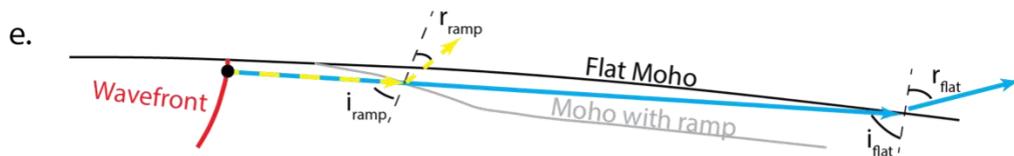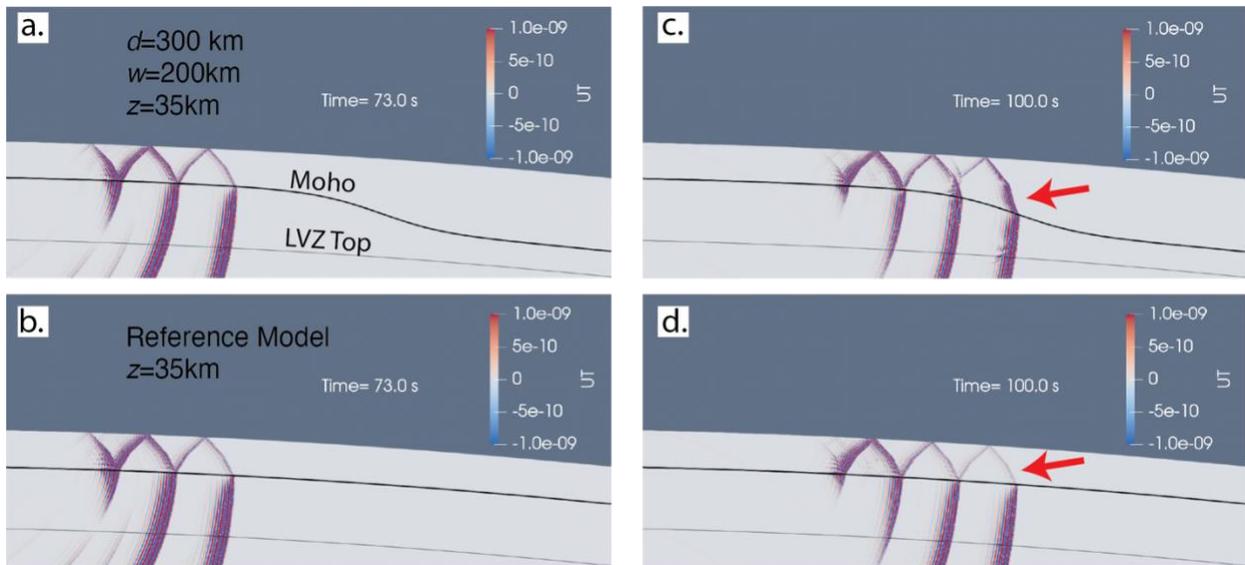
1783

1784 Fig. 7

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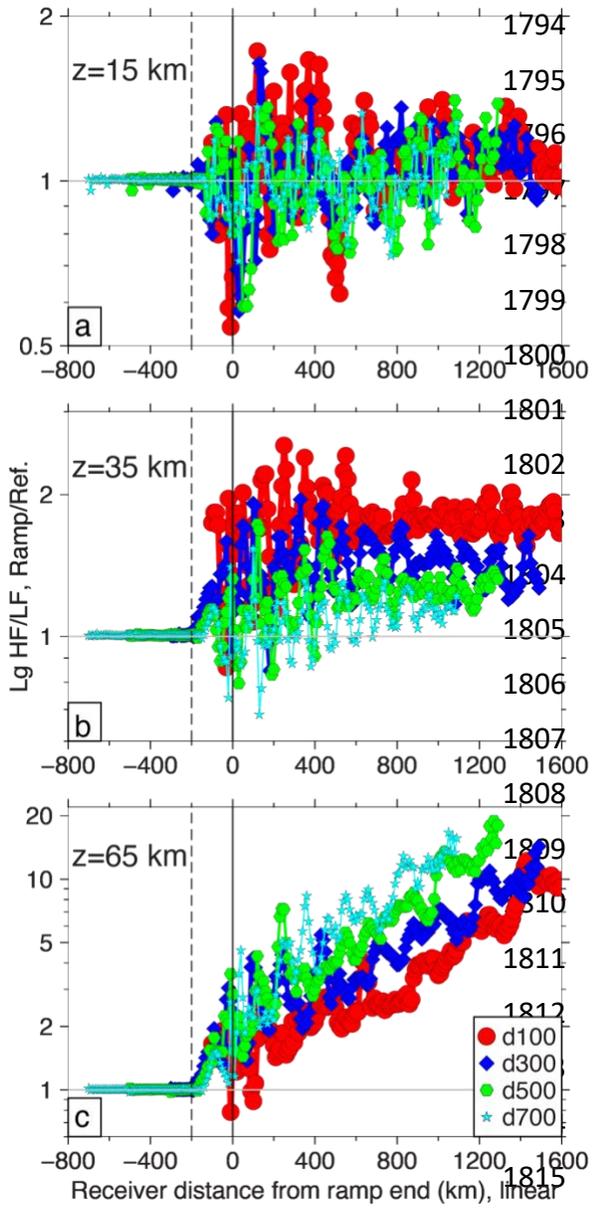
1816   Fig. 8
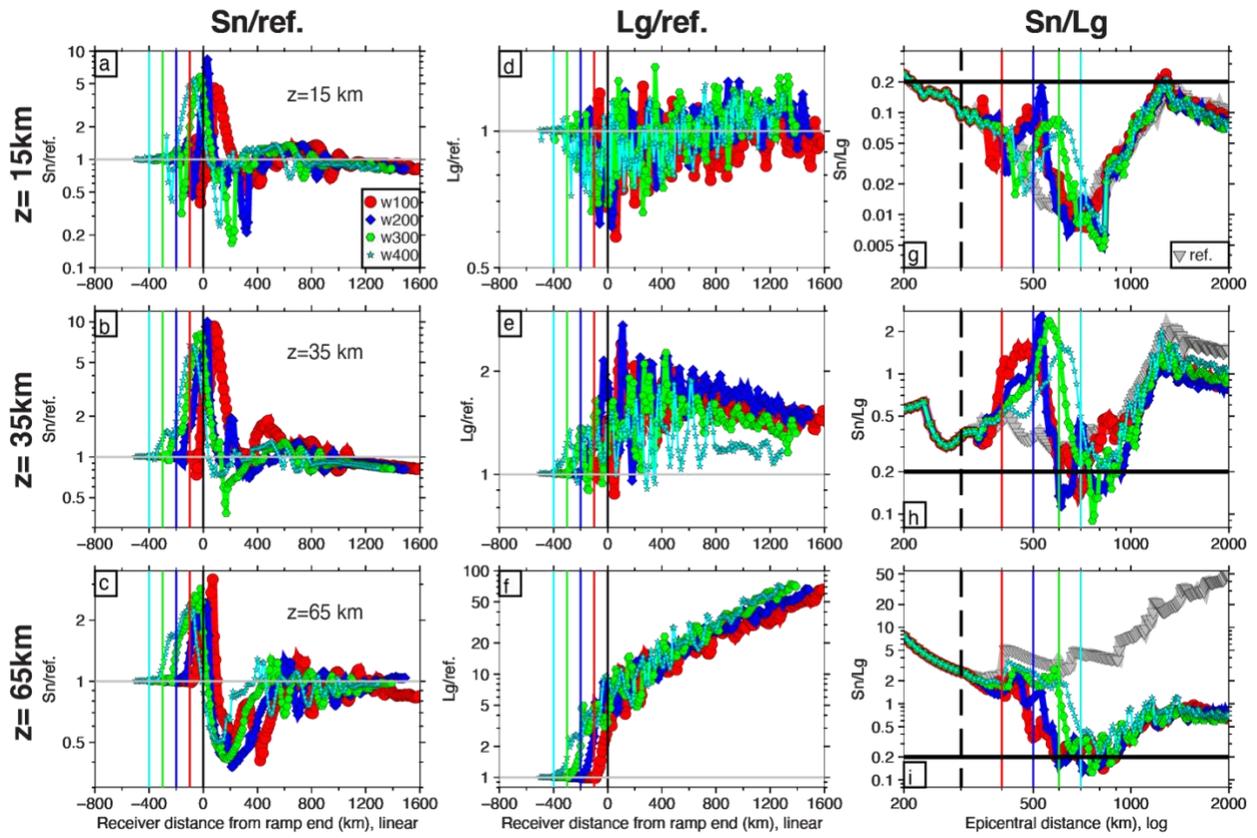
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1825   Fig. 9

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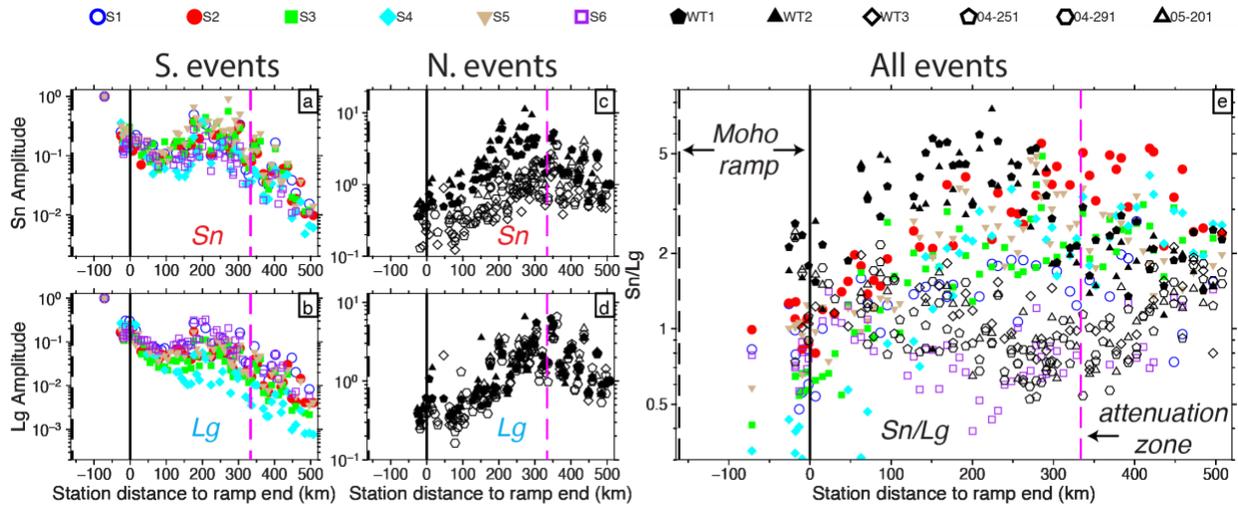
1842   Fig. 10

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South   North

Epicentral distance (km)
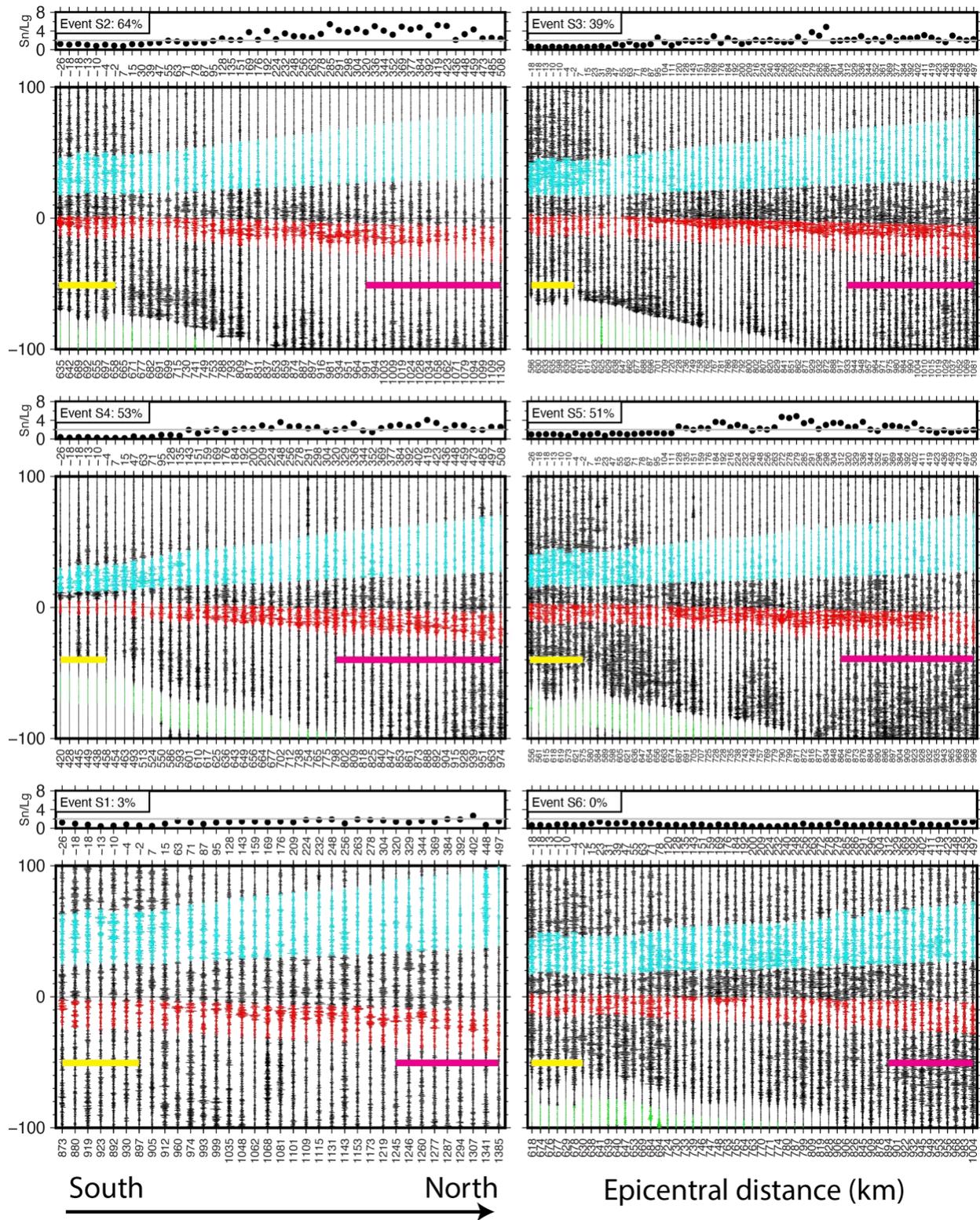
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1864   Fig. 11

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**S. events**   **N. events**   **Gangdese events**
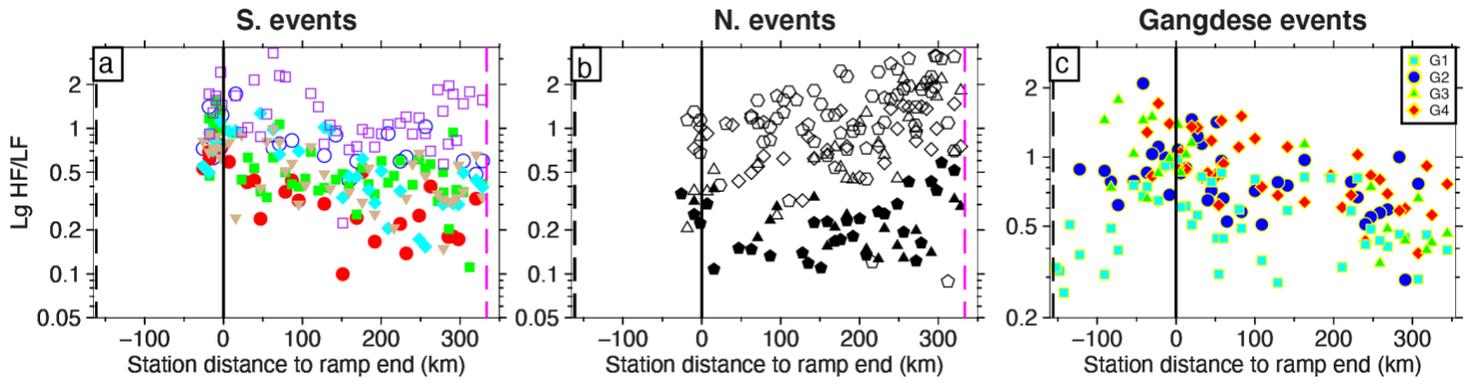
1868   Fig. 12

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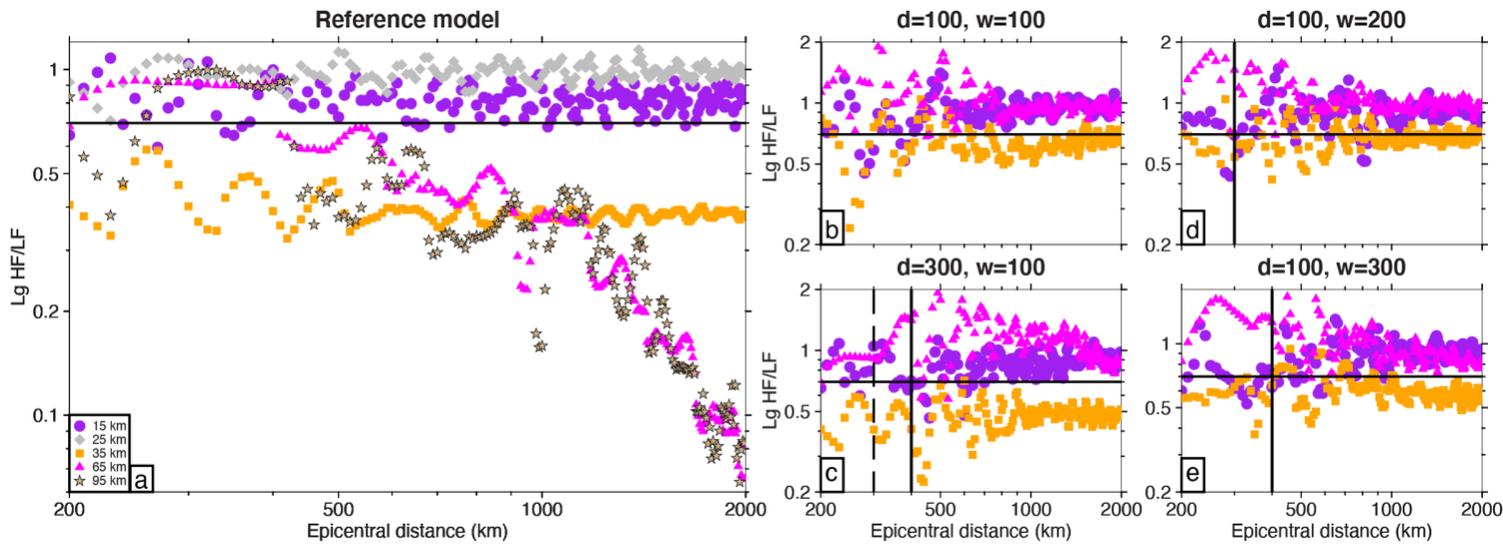
1892   Fig. 13

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