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Using teleseismic *P*-wave arrivals to calibrate the clock drift of autonomous underwater hydrophones

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ABSTRACT

1
2 Networks of Autonomous Underwater Hydrophones (AUHs) are successfully employed for
3 monitoring the low-level seismicity of mid-oceanic ridges by detecting hydroacoustic phases
4 known as *T*-waves. For a precise localization of a seismic event from *T*-wave arrival times,
5 all AUHs must be synchronized. To this effect, at the beginning of the experiment all in-
6 strument clocks are set to GPS time, which serves as a common reference. However, during
7 the experiment the instrument clock often deviates from GPS time, and since the amount
8 of deviation differs from one instrument to another, the synchronization of the AUHs dete-
9 riorates as the experiment progresses in time. Just after the instrument recovery, the time
10 difference (called “skew”) between the instrument and the GPS clocks is measured. Assuming
11 that the skew varies linearly with time, the correction of a time series for the clock drift is
12 a straightforward procedure. When the final skew cannot be determined, correcting for the
13 clock drift is not possible and any event localization becomes problematic. In this paper, we
14 demonstrate that the clock drift rate (assumed to be time-independent) can be successfully
15 estimated from arrival times of teleseismic *P*-waves, commonly recorded by AUHs. Using a
16 ray-tracing code, and accounting for the uncertainties in event hypocenter locations, origin
17 times and the Earth seismic-velocity model, confidence intervals of the estimated drift rates
18 are deduced. The validity of the approach is tested on data from two AUHs with known
19 clock-drifts. Our results show that a reliable estimation is possible for skews as small as 4 s
20 per two years (corresponding to a drift rate of about $5.5 \text{ ms} \cdot \text{day}^{-1}$). This method can also
21 be applied to correct data of other recording instruments subject to internal-clock drift, such
22 as ocean bottom seismometers, when the skew is unknown.

23 **Key words:** Autonomous underwater hydrophone, clock drift, teleseismic *P*-wave

24 INTRODUCTION

25 As demonstrated by several studies, a network of Autonomous Underwater Hydrophones (AUHs)
26 is an efficient tool to monitor the seismicity of mid-oceanic ridges (*e.g.* Fox et al. 2001; Smith
27 et al. 2002; Simão et al. 2010; Goslin et al. 2012; Tsang-Hin-Sun et al. 2016). The seismic activity
28 of spreading ridges consists of many low-magnitude earthquakes most of which are not detected
29 by land-based seismic networks (*e.g.* Bergman and Solomon 1990; Bohnenstiehl et al. 2002).
30 However, these weak events are capable of producing low-frequency acoustic waves (5–40 Hz),
31 called “Tertiary” or *T*-waves, of an appreciable amplitude (Williams et al. 2006; Okal 2008).
32 The Sound Fixing and Ranging (SOFAR) channel, acting as a waveguide, allows *T*-waves to
33 propagate in the ocean over long distances with very little attenuation. For the most efficient
34 detection of *T*-waves, an AUH is typically positioned at the depth of the SOFAR channel axis,
35 where it continuously records acoustic pressure variation. The Laboratoire Geosciences Ocean
36 (LGO) currently maintains two AUH networks, HYDROMOMAR and OHASISBIO, deployed
37 in the Atlantic and Indian oceans, respectively (Perrot 2010; Royer 2009; Tsang-Hin-Sun et al.
38 2016; Giusti et al. 2018). *T*-wave arrivals recorded by several AUHs are subsequently used to
39 deduce the location and the origin time of their source events. For a precise event localization,
40 it is crucial that the time series of all instruments are synchronized. At the beginning of an
41 experiment, the synchronization is ensured by synchronizing each AUH clock (which, in our
42 case, is based on a Temperature-Compensated Crystal Oscillator) with the GPS time shortly
43 before the instrument deployment. Once deployed at depth, the instrument clock triggers the
44 data sampling and keeps track of the timing of the acoustic records. The instrument clock is
45 however subject to a drift; in addition, drift rates vary from one instrument to another. As the
46 experiment progresses in time, the synchronization between instruments is lost. Typically, AUHs
47 are deployed for long time periods varying from one to two years and the skew can sometimes
48 reach a substantial value. As soon as possible after a recovery of an instrument, the skew between
49 the AUH and GPS clocks is measured with a microsecond precision, which allows us to correct

50 each time series and thus restore the synchronization of the hydrophone array. The correction
51 is done by assuming the skew to be a linear function of time (or equivalently, that the drift
52 rate is time-independent). This plausible assumption is indirectly confirmed by the success of
53 event localizations in all our hydroacoustic experiments. The same approach for the clock-drift
54 correction is routinely used in experiments involving data acquisition with networks of ocean
55 bottom seismometers (*e.g.* Geissler et al. 2010).

56
57 However, in some cases measuring the final skew is not possible. For example, in July 2016,
58 all five AUHs of the HYDROMOMAR network stopped functioning before they were recovered
59 (due to logistical issues, the instruments were deployed longer than expected which lead to a
60 premature exhaustion of the batteries). Without a clock-drift correction, event localization from
61 T -waves becomes too imprecise to be useful for a seismicity analysis. Several authors showed
62 that the clock drift can be estimated from interstation cross-correlations of the ambient noise
63 (*e.g.* Stehly et al. 2007; Hannemann et al. 2013; Gouédard et al. 2014; Hable et al. 2018). In a dif-
64 ferent approach, we propose to estimate the clock drift by using time markers naturally provided
65 by the Earth: teleseismic P -wave arrivals detected by AUHs (see Table 1). Acoustic signals,
66 generated by teleseismic P -waves refracting at the ocean bottom-water interface, are commonly
67 observed (*e.g.* Slack et al. 1999; Dziak et al. 2004; Simons et al. 2009; Sukhovich et al. 2015). In
68 general, source events for such signals can be easily identified with the help of land-based seismic
69 catalogs such as the one provided by the National Earthquake Information Center (NEIC, see
70 Data and Resources Section), which we have used in this study. Using a ray propagation code,
71 one can calculate the travel time between the hypocenter and the foot of the AUH mooring.
72 By adding the time that an acoustic wave takes to reach the AUH, the P -wave arrival time
73 according to the GPS clock can be predicted. Finally, the drift rate of the AUH clock can be
74 estimated by comparing expected and observed arrival times of many P -waves, which, ideally,
75 are uniformly distributed over the entire duration of the experiment. Teleseismic P -waves (*i.e.*
76 P -waves produced by distant events located at $\Delta > 50^\circ$ away from the AUH), as compared to

77 seismic waves produced by nearby events, are better suited for such analysis since they traverse
78 larger distances inside the globe, and the effect of inhomogeneities, responsible for deviation of
79 actual travel times from calculated ones, is averaged out and becomes less important. Under
80 the assumption of a time-independent drift rate, the difference between expected *P*-wave arrival
81 times and those observed in the AUH record before correction should be a linear function of time.
82 To validate the method, we applied it to data from two AUHs of the HYDROMOMAR network,
83 namely M2 and M7 (Figure 1), with known (measured) final skews. Both instruments operated
84 for two years and were recovered in June 2018. The M7 hydrophone had an unusually high clock
85 drift with a final skew of about 28 s, while that of the M2 hydrophone was approximately 7 times
86 smaller (about 3.8 s). Thus, the analysis of the M2 data also tests the sensitivity limit of the
87 method. Uncertainties in the predicted *P*-wave arrival times are estimated by a direct numerical
88 approach and take into account uncertainties in the hypocenter depths, epicenter locations and
89 origin times of the source events as well as the uncertainties in the Earth seismic-velocity model
90 (used by the ray-tracing code). For both AUHs, we find that the actual drift rate (deduced from
91 the measured skew) is located within the 95% confidence interval of the estimated drift rate. We
92 also investigate the method performance as a function of the dataset size by considering subsets
93 of *P*-wave arrivals to represent a possible scatter of estimated values. Our results demonstrate
94 that, provided that a sufficient number of teleseismic *P*-wave arrivals are available, the method
95 can reliably discern clock drifts of as small as 2 s per year.

96

97 In what follows, we first present the mathematical details of the method and then discuss
98 the results.

99

100 **METHOD**101 **Fitting**

102 The term “clock drift” designates the deviation of a given clock relative to some high-precision
 103 reference clock, whose own deviation from the absolute time can be regarded as negligibly small.
 104 In our case, we consider the drift of the AUH clock relative the GPS clock during the data
 105 acquisition period. At the beginning of the experiment the two clocks are synchronized, and the
 106 time difference (or “skew”) Δt between the time readings of two clocks is thus zero. We indicate
 107 by t_{AUH} and t_{GPS} the time intervals elapsed since the synchronization instance according to the
 108 AUH and the GPS clocks, respectively. At any moment of the record $\Delta t = t_{\text{AUH}} - t_{\text{GPS}}$. Assuming
 109 that the skew is a linear function of time, it can be expressed as :

$$\Delta t = \gamma' t_{\text{GPS}} \quad (1)$$

111
 112
 113 where γ' indicates a constant drift rate (note that at the moment of synchronization, $t_{\text{AUH}} =$
 114 $t_{\text{GPS}} = 0$ and $\Delta t = 0$). γ' can be obtained from the final skew Δt_f measured at the end of the
 115 experiment :

$$\gamma' = \frac{\Delta t_f}{t_{f,\text{GPS}}} \quad (2)$$

117
 118
 119 Since $\Delta t_f \ll t_{f,\text{GPS}}$, γ' is frequently expressed in either parts per million (ppm) or $\text{ms} \cdot \text{day}^{-1}$.

120 One can rewrite (1) to describe the clock drift in terms of t_{AUH} and t_{GPS} :

121

$$\gamma = \frac{t_{\text{AUH}}}{t_{\text{GPS}}} \quad (3)$$

122

123

124 where

125

$$\gamma = \gamma' + 1 \quad (4)$$

126

127

128 In the absence of clock drift, $\gamma = 1$ and $\gamma' = 0$. If the instrument's clock runs faster or slower
 129 than the GPS clock, one finds $\gamma > 1$ ($\gamma' > 0$) or $\gamma < 1$ ($\gamma' < 0$), respectively.

130

131 Suppose that a *P*-wave, produced by a known event, is observed in the AUH record at some
 132 moment $t_{\text{AUH}} = t_{\text{obs}}$. At the same time, using a global seismic-velocity model, one can predict
 133 the time t_{pred} at which the *P*-wave should have appeared in the record according to the GPS
 134 clock :

135

$$t_{\text{pred}} = t_{\text{or}} + t_{\text{earth}} + t_{\text{water}} \quad (5)$$

136

137

138 where t_{or} is the GPS time of the event origin (here, the time elapsed between the synchronization

instance and the event origin), t_{earth} is the P -wave travel time from the event hypocenter to the point on the ocean bottom at which the AUH mooring anchor is located, and t_{water} is the time for a refracted acoustic wave to propagate from the ocean bottom to the hydrophone (note that the design of our moorings prevents the direct transmission of P -waves or their propagation through the mooring line).

If the AUH clock drifts, t_{obs} and t_{pred} will not be the same, and based on equation (3), their ratio will yield the value of γ :

$$\gamma = \frac{t_{\text{obs}}}{t_{\text{pred}}} \quad (6)$$

and subsequently, that of γ' . Since both t_{obs} and t_{pred} are subject to uncertainties, a single P -wave arrival is not enough to reliably estimate γ . One needs to consider a sufficiently large number of P -wave arrivals, ideally distributed over the entire duration of the hydrophone record, and perform an optimization procedure. We estimate γ using a least-squares fit, which, in general, minimizes the quantity :

$$\chi^2 = \sum_{i=1}^N \frac{(y'_i - y_i)^2}{\sigma_i^2} \quad (7)$$

where y'_i are experimental data points of some variable y , y_i are its predicted values, N is the number of data points, and σ_i is an uncertainty of a data point y'_i (Bevington and Robinson 2003). The predictions y_i are obtained from some assumed functional dependence of y on an

161 independent variable x and, in general, K fitting parameters α_k :

162

$$y_i = f(\alpha_k, x_i) \quad (k = 1, \dots, K)$$

163

164

165 To estimate the “goodness-of-fit”, we use the quantity χ_ν^2 :

$$\chi_\nu^2 = \frac{\chi^2}{\nu} = \frac{\chi^2}{N - K} \quad (8)$$

166

167

168 classically known as “reduced chi square” with ν being the number of degrees of freedom (Bev-
169 ington and Robinson 2003).

170

171 Our goal is to modify the time axis of the AUH clock so that all observed arrival times
172 $t_{i,obs}$ agree with predicted arrival times $t_{i,pred}$ in the least-squares sense. Using equation (6), the
173 functional dependence of t_{pred} on t_{obs} is :

174

$$f(\gamma, t_{obs}) = \frac{t_{obs}}{\gamma} \quad (9)$$

175

176

177 This dependence is that of a straight line with a slope $b = 1/\gamma$ and passing through the
178 origin (remember that at the beginning of the experiment the instrument clock is synchronized
179 with the GPS clock and thus, necessarily, $t_{GPS} = 0$ when $t_{AUH} = 0$). It is important to note
180 that predicted arrival times $t_{i,pred}$ are based on a global seismic-velocity model which does not

181 necessarily match identically the actual seismic-velocity profile at the hydrophone site. To take
 182 into account the site effect and to estimate correctly the slope b , a common time correction δt
 183 should be applied to all predicted arrival times : $t_{i,pred} \rightarrow t_{i,pred} + \delta t$ (this correction can be either
 184 positive or negative, depending on the differences between the local and global velocity profiles).
 185 The value of δt is *a priori* unknown but can be estimated simultaneously with the slope b during
 186 the same fitting procedure from the intercept of the best-fitting line. Identifying $f(\gamma, t_{obs})$ as y'_i
 187 and $t_{i,pred} + \delta t$ as y_i , equation (7) becomes :

$$\chi^2 = \sum_{i=1}^N \frac{\left(b t_{i,obs} + a - t_{i,pred}\right)^2}{\sigma_i^2} \quad (10)$$

189
 190
 191 where $a = -\delta t$ and N is the number of observed P -wave arrivals. Equation (10) represents a
 192 linear least-squares fit with two fitting parameters. It should be noted that the overall offset a
 193 can be a sum of several offsets, as will be shown in the Results section. Both γ and γ' can be
 194 readily obtained from b (equation (4)).

195
 196 The travel time t_{water} was estimated from the bottom-hydrophone distance and the vertical
 197 sound-speed profile. In this work, we systematically took the length of the mooring line d_l as
 198 the bottom-hydrophone distance, effectively assuming that the AUH was suspended in the water
 199 column directly above the mooring anchor. The validity of this assumption is discussed at the end
 200 of this section. Taking into account that both hydrophones were deployed in flat areas (within
 201 a radius of at least 5 km around the target position, see Figure 1) and the important acoustic
 202 impedance contrast between the sea-water and the oceanic crust, the seismic waves are expected
 203 to refract into the water column at angles very close to normal. Assuming also that the sound
 204 speed varies only with depth and not laterally, the acoustic waves are also expected to propagate

205 over nearly vertical paths. The precise calculation of t_{water} in case of a depth-dependent sound
 206 speed involves an integration over the sound wave propagation path. We found however that
 207 due to the short propagation distance, with a good degree of precision, one can use a simplified
 208 expression $t_{\text{water}} = \frac{d_l}{c_{\text{mean}}}$, where c_{mean} is taken as the average of the two extreme values of
 209 the sound-speed profile (values at the ocean bottom and the SOFAR channel axis, where the
 210 AUH is positioned). By the same reason, equally well can be neglected a possible uncertainty
 211 in the acoustic travel time introduced by the uncertainty in the sound-speed profile. The time
 212 t_{water} is thus treated as a constant with no associated uncertainty. To calculate t_{water} , we used
 213 sound-speed profiles from the Generalized Digital Environmental Model produced by US Naval
 214 Oceanographic Office (Countryman and Carron 1995).

215 Estimation of uncertainties

216 According to equation (10), the calculation of χ^2 requires a knowledge of the uncertainty σ_i for
 217 each fitted point (in what follows, index i is omitted to simplify the notation). Both t_{obs} and
 218 t_{pred} are subject to its own error, which we discuss below.

219

220 *Uncertainty in observed arrival times*

221 The uncertainty in t_{obs} is user-estimated and represents the precision with which an operator
 222 performing the time-picking is able to identify the onset of a *P*-wave. It is not a true statis-
 223 tical error but rather a numerical criterion of the quality of the time-pick. We thus call this
 224 uncertainty “Quality Criterion” (*QC*). Obviously, its value mostly depends on the complexity
 225 of the signal onset and the signal-to-noise ratio, but also on the operator’s experience. For the
 226 most complicated *P*-wave arrivals, *QC* may amount to 2 s. When the signal onset is clear and
 227 the time-pick is evident, we ascribed to *QC* a value of 0.05 s (equivalent to approximately 12
 228 sampling intervals). Two examples of seismic phases with low and high *QC* are presented in
 229 Figure 2. As explained later, the *QC* was used to make a decision whether a given arrival should

230 or should not be included in the fit. In the fit itself, no uncertainties were ascribed to t_{obs} .

231

232 *Uncertainty in predicted arrival times*

233 The predicted arrival time t_{pred} is a sum of the event origin time, the P -wave travel time in the
 234 Earth and the acoustic-wave travel time in the water column (equation (5)). The uncertainty
 235 σ_{th} in t_{pred} is thus a combination of the uncertainties in the event origin time σ_{or} and the travel
 236 time σ_{prop} . Its calculation is presented after the discussion of σ_{or} and σ_{prop} .

237

238 Assuming an event with a hypocenter located at depth d_0 , and epicenter's latitude and lon-
 239 gitude θ_0 and ϕ_0 , respectively, several factors contribute to the σ_{prop} :

240

- 241 – the uncertainty in the focal depth σ_d , which characterizes the spread around the point d_0 of
 242 the possible depths of the hypocenter along the radial line passing through the point (θ_0, ϕ_0) ;
- 243 – the uncertainty in the hypocenter position σ_p , which, at any given hypocenter depth, char-
 244 acterizes the spread around the point (θ_0, ϕ_0) of possible hypocenter positions on the surface of
 245 a sphere whose radius is equal to the Earth radius minus the hypocenter depth ;
- 246 – the uncertainty in the Earth seismic-velocity model σ_{vm} .

247

248 Due to the complexity of travel-time calculations, an analytical derivation of an error prop-
 249 agation formula to calculate σ_{prop} directly from the above uncertainties is a complicated task.
 250 Instead, we resorted to a simpler numerical approach. We first generated a collection of points Π ,
 251 randomly distributed around the point (d_0, θ_0, ϕ_0) , each point representing a possible hypocenter
 252 position. Second, we generated an ensemble of velocity models randomly distributed around the
 253 reference model (ak135 in our case) to reflect the uncertainty in our knowledge of the seismic

254 velocities in the Earth. For each pair “point in Π /velocity model”, we then calculated a travel
 255 time to the AUH. This resulted in an ensemble of travel times t_{tr} , from which the uncertainty
 256 σ_{prop} was estimated.

257
 258 In addition to the event’s d_0 , θ_0 and ϕ_0 , the NEIC seismic catalog generally reports two er-
 259 rors, named *depthError* and *horizontalError* (both measured in km) and defined as the largest
 260 projections of the three principal axes of the error ellipsoid on a vertical axis and a horizontal
 261 plane, respectively. We used the value of *depthError* as σ_d to generate an ensemble of points D
 262 which are randomly distributed according to the normal law with the mean d_0 and the standard
 263 deviation σ_d .

264
 265 The uncertainty σ_p was assumed to be depth-independent and equal to *horizontalError*.
 266 For each point d_j in D (j is a dummy index indicating a particular depth value), the haversine
 267 formula was used to make a transition from σ_p to uncertainties in latitude $\sigma_{lat,j}$ and longitude
 268 $\sigma_{lon,j}$. Next, for each d_j , a random ensemble P_j was generated, with θ and ϕ distributed around θ_0
 269 and ϕ_0 according to the normal law with standard deviations $\sigma_{lat,j}$ and $\sigma_{lon,j}$, respectively. Points
 270 P_j , combined for all depths d_j , form the collection of points Π . In a more rigorous approach to
 271 generate the collection Π , one might attempt taking into account the directivity of the error
 272 ellipse axes (not reported by the NEIC catalog). However, the simpler approach adopted in this
 273 paper is sufficient to demonstrate the effectiveness of the method.

274
 275 As an example, Figure 3 illustrates the procedure of estimating the spread in angular dis-
 276 tance Δ between the M2 hydrophone and an event, which happened on September 8, 2017 at
 277 04:49:19.18 UTC (Table 1). According to the NEIC catalog, the event’s depth d_0 was 47.39 km
 278 and its epicenter, located at $\theta_0 = 15.02^\circ$ N and $\phi_0 = 93.90^\circ$ W, was separated by the angular
 279 distance Δ_0 of 57.35° from the hydrophone location. The catalog reports the values of 6.2 and
 280 3.7 km for the *horizontalError* and *depthError*, respectively. A histogram of the corresponding

ensemble D is shown in Figure 3(a). Figures 3(b) and 3(c) show histograms of latitudes and longitudes of points in Π (all depths combined), respectively. Finally, Figure 3(d) displays a histogram of angular distances Δ separating each point in Π and the mooring anchor of the M2 hydrophone. For each event, ensemble D contained 100 points, while each ensemble P_j contained 900 points. All depths combined, each event's hypocenter was represented by a collection of 90000 points.

Calculating the travel times from all points in Π to the AUH location would result in a collection of travel times distributed with some standard deviation σ_{prop} which is a function of σ_d and σ_p . The remaining source of uncertainty contributing to the σ_{prop} is due to the uncertainty σ_{vm} in the velocity model used for calculation of travel times. To take into account σ_{vm} , a simple and straightforward way is to calculate travel times between points in Π and the AUH location using many different velocity models. An ensemble of such velocity models should be representative of the uncertainty in the principal velocity model and its realization is discussed later. This will result in an even larger collection of t_{earth} , whose standard deviation will be a function of all three uncertainties σ_d , σ_p and σ_{vm} .

Travel time calculations were performed with the *ObsPy* Python toolbox (see Data and Resources Section). As a principal model, we used the ak135 model introduced by Kennett et al. (1995). According to these authors, with the exception of the inner core and the D'' region, the P -wave speed is constrained within ± 0.01 km/s with a reasonable level of confidence. Upon examination of Figure 9 in Kennett et al. (1995), it seems that one can assign a higher uncertainty of ± 0.015 km/s to the speed of P -waves in both the inner core and the D'' region. We generated 265 velocity models uniformly distributed around ak135 within the specified limits. Another requirement stipulated that all velocity gradients in any generated velocity model did not deviate by more than 35% of the corresponding gradients in ak135. Figure 4 compares our velocity models with the ak135 model. Using as an example the event presented in Figure 3,

we compare in Figure 5 the effect on the resulting collection of travel times when taking into account : (a) the uncertainty in the hypocenter location only, (b) the uncertainty in the velocity model only, (c) both of these uncertainties. Vertical dotted lines indicate the travel time predicted for this event by the ak135 model. To reliably estimate the standard deviation of a resulting travel time distribution, we used a bootstrapping procedure. The standard deviations of a large number of subsets, randomly drawn from the travel time distribution, were calculated (in all three cases we have taken 10000 subsets, each composed of 40000 values in case of the distributions shown in Figures 5(a) and (e) and 80 values in case of the distribution shown in Figure 5(c)). The histograms of the resulting collections of standard deviations are shown in Figures 5(b),(d) and (f). The mean of each collection is taken as the final value of σ_{prop} while the number of significant decimal places in σ_{prop} is estimated from the standard deviation of each collection. Comparison of Figures 5(b) and (d) shows that for this event the effect of the uncertainty in the velocity models on the width of the travel time distribution is significantly smaller than the effect of the uncertainty in the hypocenter location. Nevertheless, taking into account both uncertainties results in a slight increase of σ_{prop} (compare the values displayed in Figures 5(b) and (f)). Moreover, as shown in Tables S1 and S2 of Supplemental material, for some events σ_{vm} can provide a significant contribution to the final error σ_{prop} . Therefore, for each event we systematically applied all three uncertainties σ_d , σ_p and σ_{vm} to estimate σ_{prop} .

In a final step, the uncertainty σ_{prop} was combined with the uncertainty in origin time σ_{or} , the latter being retrieved from the catalog of the International Seismological Centre (ISC, see Data and Resources Section) for the selected NEIC solutions. For all events, we found that the resulting distributions of travel times follow closely the normal law (see Figure 5(e) showing the September 8, 2017 event as an example). We assumed the event origin time t_{or} to be also normally distributed with the variance σ_{or}^2 . The sum of two normal random variables of variances σ_1^2 and σ_2^2 also being a normal random variable of variance σ^2 equal to the sum of both variances (see for example Mood et al. (1974)), the error in an event origin time can be accounted for by

335 considering $\sigma_{th} = \sqrt{\sigma_{prop}^2 + \sigma_{or}^2}$.

336
 337 Tables 2 and 3 report the values of QC and all uncertainties for each observed P -wave arrival.
 338 Arrivals for which QC was larger than the corresponding σ_{th} were rejected and not used in the
 339 fit. For the remaining arrivals, σ_{th} was used as σ in equation (10).

340
 341 In this work, the location of the AUH corresponded to that of the mooring anchor whose
 342 location was known within a few meters. The bottom-hydrophone distance was taken to be equal
 343 to the length of the mooring line. It may happen that, in presence of deep oceanic currents, the
 344 mooring line will slant in the current direction and thus both AUH elevation and position will be
 345 different from their assumed values. A quantitative estimation of vertical and horizontal offsets
 346 introduced by currents is a complicated task which should be performed separately for each
 347 hydrophone location since in general the currents vary from one site to another. Moreover, in a
 348 given area, the offsets most likely vary with time as the buoy carrying the hydrophone moves
 349 around some equilibrium position. In a similar deployment of our AUH in the Indian Ocean, we
 350 were able to measure a horizontal offset of approximately 500 m for a 3560 m long line which is
 351 equivalent to a slant angle of about 8° . The M2 and M7 hydrophones had shorter mooring lines,
 352 2750 and 2450 m, respectively. Using simple geometric arguments, for the same slant angle, we
 353 estimated for the M2 and M7 hydrophones a horizontal offset of about 380 and 340 m and a
 354 vertical offset of about 27 and 24 m, respectively. The mean speed of sound c_{mean} being about
 355 $1508 \text{ m} \cdot \text{s}^{-1}$ at both sites, the error in t_{water} induced by neglecting the vertical offset would be
 356 less than 0.02 s, one order of magnitude smaller than the smallest estimated uncertainty σ_{th} .
 357 Similarly, the horizontal error in the location of the AUH ($< 400 \text{ m}$) should not affect signifi-
 358 cantly the value of σ_{th} compared to the combined effect of the uncertainties σ_d and σ_p . Therefore,
 359 neglecting possible depth and position uncertainties in the hydrophone position seemed to be
 360 justified.

362 RESULTS

363 Using the NEIC seismic catalog, we first identified events of magnitude 6 or higher that we
 364 expected to be strong enough to produce acoustic signals detectable by the HYDROMOMAR
 365 network. Next, the origin times of the events, listed by the catalog in UTC time, were corrected
 366 for the leap second offset to be expressed in GPS time. Using the *ObsPy* toolbox, arrival times
 367 of all possible seismic phases were predicted and then matched with the acoustic signals (when
 368 observed). For the best clock-drift estimation, only signals with sufficiently high signal-to-noise
 369 ratio were retained. It is worth noting that not all of the selected events produced useful acous-
 370 tic signals in the time series of both instruments (possibly, due to site effects and/or different
 371 ambient noise levels at the hydrophone locations on the day of detection). Jointly, the M2 and
 372 M7 hydrophones recorded 27 events, listed in Table 1 and shown in Figure 6.

373
 374 Tables 2 and 3 provide more details on the detected events, including the seismic phases
 375 expected to arrive at the hydrophone locations. Seismic phase names follow the current *ObsPy*
 376 convention according to which *PKPab* and *PKPbc* are both labeled as *PKP* while *PKPdf* is
 377 labeled as *PKIKP*. In case of triplications, it was not obvious to establish an unambiguous cor-
 378 respondence between an observed acoustic signal and its corresponding seismic phase for the
 379 phases predicted to arrive close in time. To avoid such ambiguity, we preferred to combine the
 380 seismic phases whose travel times were separated by less than 0.5 s, to produce a joint collection
 381 of travel times. For example, for the July 29, 2016 event, the combined phases were *PKIKP* and
 382 *PKiKP*, while for the August 31, 2016 and September 24, 2016 events, two *PKP* phases were
 383 combined (Table 2). For these events, the arrival time t_{pred} was taken as the average of the arrival
 384 times predicted with the ak135 model for each of the combined phases.

385
 386 In the ak135 model, the Earth is approximated by a solid sphere with an uppermost 20-km-
 387 thick layer in which the *P*-wave velocity is equal to $5.8 \text{ km} \cdot \text{s}^{-1}$. In all travel-time calculations,

388 ray paths originated at the earthquake hypocenter and ended at the surface of the sphere, which
 389 in our case corresponds to the ocean surface. Thus, all travel times calculated with ak135 for
 390 the P -waves arriving at the hydrophone site will exceed the actual travel times by the amount
 391 approximately equal to the ratio of the water depth at the hydrophone site to the P -wave ve-
 392 locity ($5.8 \text{ km} \cdot \text{s}^{-1}$). The exact value of this offset obviously depends on the ray path and may
 393 vary slightly from one event to another, but to a good degree of precision can be considered the
 394 same for all arrivals. The presence of this time offset, as well as any eventual offset due to a site
 395 effect, is automatically taken into account by the parameter a in fit (10).

396
 397 For the M7 hydrophone, the measured clock drift γ'_m was found to be of 0.437 ppm or
 398 $37.72 \text{ ms} \cdot \text{day}^{-1}$ (the corresponding final skew Δt_f , rounded to two decimal places, is 28.10 s;
 399 note that its actual value is known with a microsecond accuracy). Figure 7(a) presents the pre-
 400 dicted t_{pred} versus observed t_{obs} arrival times. At the scale of the entire experiment, the effect
 401 of the observed clock drift is not discernible, as illustrated by the dashed line of slope 1 rep-
 402 resenting the case of no clock-drift (i.e. $t_{pred} = t_{obs}$). To make the presence of the clock drift
 403 evident, Figure 7(b) displays the differences ($t_{obs} - t_{pred}$) versus the event number. Since γ'_m is
 404 positive, one expects that for each event $t_{obs} > t_{pred}$ and their difference should be increasing
 405 linearly as the experiment progresses in time (equation (1)). This is indeed the case as can be
 406 seen in Figure 7(b). By fitting the data shown in Figure 7(a) (with the exception of two ex-
 407 cluded points not consistent with the QC criterion), the estimated drift rate γ'_e was found to
 408 be 0.467 ppm (fit M7-1). The estimated γ'_e and observed γ'_m are in good agreement, the latter
 409 falling within 95% confidence interval of the former. A reasonably low value of χ^2_ν of 1.49 is
 410 consistent with the hypothesis of a constant drift rate. The value of the parameter a is 0.52 s
 411 and agrees well with a quick estimate of 0.59 s, calculated as a ratio of the water depth at the
 412 M7 site (3.4 km) to the P -wave velocity (5.8 km/s) assumed by ak135 in the corresponding layer.

413
 414 To test the sensitivity limit of the method, we performed a similar analysis on the data

415 from the M2 hydrophone, for which the skew measured at the end of the experiment was al-
 416 most 10 times smaller than that of the M7 hydrophone. More precisely, γ'_m was found to be
 417 0.059 ppm or $5.12 \text{ ms} \cdot \text{day}^{-1}$ (the corresponding Δt_f , rounded to two decimal places, is 3.82 s).
 418 Figures 8(a) and 8(b) present t_{pred} versus t_{obs} arrival times and their differences, respectively.
 419 Two points inconsistent with the *QC* criterion (Table 2), were not used in the first fit indicated
 420 as fit M2-1. The fit yields $\gamma'_e = 0.083$ ppm and a 95% confidence interval bracketing γ'_m . The
 421 consistency between a linear model and the observed data is again supported by a χ^2_ν of 0.41.
 422 The parameter a is equal to 0.57 s, again, consistent with a quick estimate of 0.62 s (the M2 site
 423 depth is 3.6 km).

424

425 The agreement between estimated and measured clock drifts can be improved even further
 426 if one takes into account that the ak135 model is developed from arrivals of seismic waves gen-
 427 erated by earthquakes mostly located in the oceans and recorded by land-based seismological
 428 stations (Kennett et al. 1995). One might expect that in our case the ak135 would best describe
 429 *P*-waves from earthquakes originating in the continental crust and detected by a receiver (AUH)
 430 located in the ocean, since this situation is somehow similar to that of Kennett et al. (1995)
 431 except that receivers (at sea) and sources (on land) are switched. There is an equivalent global
 432 seismic-velocity model called ak135-f whose version for an oceanic lithosphere (see Data and
 433 Resources Section) mainly differs from the ak135 model only in the range between 0 and 35 km
 434 (water layer, sediment layer, oceanic crust and upper mantle instead of a 35-km-thick conti-
 435 nental crust in ak135). To make sure that ak135 is applicable to all events used in the fit, we
 436 excluded, in fits M2-2 and M7-2, all oceanic events (i.e. events whose epicenters are in ocean)
 437 shallower than 35 km (Table 1), since they certainly did not occur in a continental crust. With
 438 these events excluded, we find γ'_e of 0.434 ppm (fit M7-2, Figure 7(c)) and 0.059 ppm (fit M2-2,
 439 Figure 8(c)) for the M7 and M2 hydrophones, respectively. The values of the offset a , -0.93 s and
 440 0.33 s for the M7 and M2 hydrophones, respectively, can be used to approximately quantify the
 441 site effects by subtracting from the estimated offset a the offset due to the ak135 model (which

amounts, as explained above, to 0.59 s and 0.62 s for the M7 and M2 hydrophones, respectively).
 One obtains the site contribution of -1.52 s for the M7 hydrophone and -0.29 s for the M2
 hydrophone. This means that at both hydrophone locations, the P -waves arrive systematically
 later as predicted by the ak135 model (remember that in (10), $a = -\delta t$). This explanation seems
 to be quite reasonable considering that both hydrophones are located in the area influenced by a
 low-velocity anomaly due to the presence of the Azores hotspot (Yang et al. 2006; Silveira et al.
 2006), with the M7 hydrophone actually positioned in the presumed hotspot location (Silveira
 et al. 2006). The amount of slowing down of P -waves is consistent with the results of Silveira
 et al. (2010) who found an evidence for the existence of a low P -wave velocity profile in the
 first 150 km of the upper mantle beneath the Azores. The site effect is less obvious in fits M7-1
 and M2-1, possibly due to the presence of the arrivals of the P -waves originating in the oceanic
 crust. Table 4 summarizes the results of all four fits. More information on each fit can be found
 in Supplemental Material.

Figure 9 is a different way to present the results of the fits. For both hydrophones, it compares
 the observed time differences ($t_{obs} - t_{pred}$) with those expected from equation (1) (after having
 taken into account the time offset a) at moments t_{pred} based on measured γ'_m and estimated γ'_e
 drift rates.

To check how well the method performs when the sample size is decreased, we investigated
 for each hydrophone the spread of γ'_{red} values obtained from fits performed on reduced sets
 containing approximately 80% and 60% of the arrivals used in fits M2-2 and M7-2. Subsets were
 formed by choosing all possible combinations of n arrivals (irrespective of their order within a
 given combination) out of a larger collection of N arrivals (the total number of combinations C
 is given by $\frac{N!}{(N-n)!n!}$). For the M2 hydrophone this resulted in selecting $n = 12$ and $n = 9$
 arrivals out of $N = 15$ arrivals and for the M7 hydrophone in selecting $n = 10$ and $n = 8$
 arrivals out of $N = 13$ arrivals. Figure 10 presents the histograms of obtained γ'_{red} values while

469 the numerical results are presented in Table 5. In all cases, the mean value γ'_{mean} remains close
 470 to γ'_e found by fitting all arrivals. To measure the scatter of γ'_{red} values, for each collection shown
 471 in Figure 10 we calculated the ratio of the standard deviation σ_n to γ'_{mean} . This quantity can be
 472 considered as an analog of the coefficient of variation σ/μ employed in statistics to characterize
 473 the dispersion of a probability distribution. Only in one case the ratio exceeds 20%, implying a
 474 reasonably narrow spreading of the estimated values γ'_{red} around the measured value γ'_m (this
 475 spread decreases when number of arrivals used in the fit increases). Finally, in each case we
 476 considered an interval containing 95% of γ'_{red} values. This interval can be viewed as an analog
 477 of a 95% confidence interval of a true random variable. As can be seen from Figure 10, for both
 478 hydrophones the measured value γ'_m always falls within the corresponding 95% interval.

479 CONCLUSION

480 We demonstrated that the drift rate of an AUH internal clock can be successfully estimated
 481 using arrivals of teleseismic *P*-waves. We employed a ray-tracing code in combination with un-
 482 certainties in the location, depth and origin time of the events as well as uncertainties in the
 483 Earth seismic-velocity model, to place precision limits on estimated drift rates. By applying
 484 the method to data from two hydrophones with known clock drifts, we found that the drift
 485 rate resulting in a skew as small as 4 s per two years (about $5.5 \text{ ms} \cdot \text{day}^{-1}$) can be estimated
 486 with a good precision. Unlike techniques involving cross-correlations of the ambient noise (*e.g.*
 487 Gouédard et al. 2014; Hable et al. 2018), our method cannot discern time variations in the drift
 488 rate, which we assumed to be constant over the entire duration of the experiment. However, the
 489 proposed approach is significantly less computationally intensive and simpler to implement. At
 490 the same time, the assumption of a time-invariant drift rate seems quite reasonable as evidenced
 491 by the low values of χ^2_ν obtained in all fits. The narrow spread of drift rates estimated from
 492 reduced subsets of arrival times demonstrates that the method is not very sensitive to the size
 493 of the dataset. In view of designing a practical tool for estimating the clock drift, selecting only

494 events occurring in the continental crust and all other events with hypocenters located deeper
495 than 35 km and applying the ak135 model provides a simple recipe and good results. A further
496 improvement could be a calculation of travel times with a model which combines a continental
497 structure on the source side, an oceanic structure on the receiver side and a common part for
498 depths > 35 km. Additionally, the uncertainties in the observed arrival times t_{obs} could be quan-
499 tified, e.g. as recently proposed by Simon et al. (2020). Our method is not limited to the field
500 of hydroacoustics, and can be applied to any autonomous recording instruments, such as ocean
501 bottom seismometers.

503 Data and resources

504 The AUH data used in this study was acquired by and belongs to the LGO. The NEIC seismic cat-
505 alog (<https://earthquake.usgs.gov/earthquakes/search/>, last accessed February 2020) was used
506 to identify the source events of the observed *P*-wave arrivals. The full list of the networks con-
507 tributing to the NEIC catalog can be found at <https://earthquake.usgs.gov/data/comcat/contributor/>
508 (last accessed February 2020). The ISC catalog (<http://www.isc.ac.uk>, last accessed February
509 2020; Bondár and Storchak 2011) was used to obtain the uncertainties on the event origin
510 times. Maps in Figures 1 and 6 were made using the Generic Mapping Tools version 4.5.18
511 (www.soest.hawaii.edu/gmt; Wessel et al. 2013). Theoretical travel times were computed using
512 the package *obs.py.taup* which is a part of a publicly available *ObsPy* tool box (<https://docs.obspy.org/packages>
513 last accessed August 2019; Beyreuther et al. 2010). The oceanic version of the ak135-f model
514 is available at <http://rse.anu.edu.au/seismology/ak135/ak135f.html> (last accessed September
515 2020). The fitting was performed using *Curve Fitting Toolbox* of MATLAB. The toolbox allows
516 one to define a model and reports both values and confidence intervals of the fitting parameters.
517 Supplemental material to this paper provides information on how, for each event, uncertainties
518 in the seismic-velocity model and in hypocenter position contribute to the final uncertainty in
519 travel time. It also includes extra details on the fits.

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Table 1: Detected teleseismic events.

Event time (UTC)	Lat. (°)	Lon. (°)	Depth (km)	Magn.	Place	Oceanic AUH
2016-07-29T21:18:24.740Z	+18.54	+145.51	196.00	7.7	29km SW of Agrihan, Northern Mariana Islands	M2,M7
2016-08-19T07:32:22.710Z	-55.29	-031.88	10.00	7.4	South Georgia Island region	✓ M7
2016-08-29T04:29:57.860Z	-0.05	-017.83	10.00	7.1	North of Ascension Island	✓ M7
2016-08-31T03:11:34.420Z	-3.68	+152.79	476.00	6.8	39km E of Namatanai, Papua New Guinea	M2
2016-09-24T21:28:41.700Z	-19.78	-178.24	596.40	6.9	107km NNE of Ndoi Island, Fiji	M2
2016-11-24T18:43:47.710Z	+11.91	-088.90	10.00	6.9	156km SSW of Puerto El Triunfo, El Salvador	✓ M2
2016-11-25T14:24:30.710Z	+39.27	+073.98	17.00	6.6	47km NE of Karakul, Tajikistan	M7
2016-12-25T14:22:27.010Z	-43.41	-073.94	38.00	7.6	41km SW of Puerto Quellon, Chile	M7
2017-01-22T04:30:22.960Z	-6.25	+155.17	135.00	7.9	35km WNW of Panguna, Papua New Guinea	M2
2017-02-21T14:09:04.320Z	-19.28	-063.90	595.98	6.5	41km E of Padilla, Bolivia	M2,M7
2017-04-03T17:40:18.560Z	-22.68	+025.16	29.00	6.5	132km WSW of Moijabana, Botswana	M2,M7
2017-04-24T21:38:30.820Z	-33.04	-072.06	28.00	6.9	40km W of Valparaiso, Chile	✓ M2,M7
2017-06-02T22:24:47.440Z	+54.03	+170.92	5.00	6.8	200km NW of Attu Station, Alaska	✓ M2
2017-06-14T07:29:04.390Z	+14.91	-092.01	93.00	6.9	2km SSW of San Pablo, Guatemala	M2,M7
2017-06-22T12:31:03.490Z	+13.72	-090.97	38.12	6.8	28km SW of Puerto San Jose, Guatemala	M2
2017-07-17T23:34:13.740Z	+54.44	+168.86	10.00	7.7	202km ESE of Nikol'skoye, Russia	✓ M2
2017-07-20T22:31:11.260Z	+36.93	+027.41	7.00	6.6	11km ENE of Kos, Greece	M2,M7
2017-09-08T04:49:19.180Z	+15.02	-093.90	47.39	8.2	101km SSW of Tres Picos, Mexico	M2,M7
2017-09-19T18:14:38.090Z	+18.55	-098.49	48.00	7.1	1km E of Ayutla, Mexico	M2,M7
2017-11-12T18:18:17.180Z	+34.91	+045.96	19.00	7.3	29km S of Halabjah, Iraq	M2,M7
2018-01-10T02:51:33.290Z	+17.48	-083.52	19.00	7.5	44km E of Great Swan Island, Honduras	✓ M2,M7
2018-01-14T09:18:45.540Z	-15.77	-074.71	39.00	7.1	38km SSW of Acari, Peru	M2,M7
2018-01-23T09:31:40.890Z	+56.00	-149.17	14.06	7.9	280km SE of Kodiak, Alaska	✓ M2,M7
2018-02-16T23:39:39.280Z	+16.39	-097.98	22.00	7.2	3km S of San Pedro Jicayan, Mexico	M2,M7
2018-03-26T09:51:00.430Z	-5.50	+151.40	40.00	6.7	139km E of Kimbe, Papua New Guinea	M2
2018-03-29T21:25:36.790Z	-5.53	+151.50	35.00	6.9	150km E of Kimbe, Papua New Guinea	✓ M2
2018-04-02T13:40:34.840Z	-20.66	-063.01	559.00	6.8	11km NNE of Carandayti, Bolivia	M2,M7

List of events (from the National Earthquake Information Center seismic catalog) which have produced usable acoustic signals in the time series of the M2 and/or M7 hydrophones. The instruments were deployed on June 17 and June 13, 2016 and recovered on July 2 and June 28, 2018, respectively. Column “Oceanic” identifies the events whose epicenters are oceanic and depths are shallower than 35 km.

Table 2: Events detected by the M2 hydrophone (39.42° N, 34.11° W, site depth : 3600 m).

	Event time (UTC)	Depth (km)	Δ (°)	Phase	Travel time (s)	σ_p (km)	σ_d (km)	σ_{or} (s)	σ_{prop} (s)	σ_{th} (s)	QC (s)	σ (s)
1	2016-07-29T21:18:24.740Z	196.00	122.34	PKIKP	1111.57	7.50	1.80	2.80	0.30	2.82	0.05	2.82
				PKiKP	1111.77							
2	2016-08-31T03:11:34.420Z	476.00	143.88	PKP	1119.28	8.10	1.90	1.27	0.35	1.32	0.05	1.32
				PKP	1119.32							
				PKIKP	1120.34							
				PKiKP	1124.00							
3	2016-09-24T21:28:41.700Z	596.40	143.51	PKP	1106.38	9.20	3.50	1.67	0.49	1.74	0.40	1.74
				PKP	1106.43							
				PKIKP	1107.55							
				PKiKP	1111.17							
4	2016-11-24T18:43:47.710Z	10.00	55.47	P	574.89	5.60	1.60	2.07	0.44	2.12	0.30	2.12
5	2017-01-22T04:30:22.960Z	135.00	145.93	PKIKP	1162.68	7.30	1.80	2.48	0.27	2.49	0.50	2.49
				PKP	1163.60							
				PKP	1164.30							
				PKiKP	1166.86							
6	2017-02-21T14:09:04.320Z	595.98	64.70	P	580.46	9.60	3.00	1.83	0.66	1.94	0.20	1.94
7	2017-04-03T17:40:18.560Z	29.00	82.92	P	742.01	7.50	1.80	2.25	0.45	2.30	0.30	2.30
8	2017-04-24T21:38:30.820Z	28.00	80.22	P	727.88	4.90	1.80	1.48	0.38	1.53	1.00	1.53
9	2017-06-02T22:24:47.440Z	5.00	84.46	P	753.68	6.70	1.70	1.84	0.43	1.89	0.70	1.89
10	2017-06-14T07:29:04.390Z	93.00	55.97	P	568.62	2.80	2.20	4.34	0.28	4.35	0.40	4.35
11	2017-06-22T12:31:03.490Z	38.12	55.91	P	574.00	6.80	3.50	2.96	0.61	3.02	0.20	3.02
12	2017-07-17T23:34:13.740Z	10.00	84.45	P	752.80	7.00	1.80	1.37	0.45	1.44	2.00	–
13	2017-07-20T22:31:11.260Z	7.00	47.60	P	516.42	4.30	1.70	1.59	0.41	1.64	2.00	–
14	2017-09-08T04:49:19.180Z	47.39	57.35	P	583.26	6.20	3.70	2.47	0.55	2.53	0.40	2.53
15	2017-09-19T18:14:38.090Z	48.00	58.82	P	593.46	4.50	1.80	2.32	0.35	2.35	0.05	2.35
16	2017-11-12T18:18:17.180Z	19.00	61.96	P	618.55	4.90	1.70	2.62	0.41	2.65	0.20	2.65
17	2018-01-10T02:51:33.290Z	19.00	47.95	P	517.26	5.80	1.70	1.65	0.49	1.72	0.30	1.72
18	2018-01-14T09:18:45.540Z	39.00	66.71	P	646.91	7.00	1.80	2.43	0.45	2.47	0.30	2.47
19	2018-01-23T09:31:40.890Z	14.06	70.19	P	672.23	6.00	2.80	1.57	0.54	1.66	0.05	1.66
20	2018-02-16T23:39:39.280Z	22.00	59.69	P	602.66	4.30	1.80	2.49	0.39	2.52	0.30	2.52
21	2018-03-26T09:51:00.430Z	40.00	145.87	PKIKP	1174.28	6.50	0.80	2.39	0.20	2.40	0.50	2.40
				PKP	1174.92							
				PKP	1175.40							

				PKiKP	1178.41							
22	2018-03-29T21:25:36.790Z	35.00	145.88	PKIKP	1174.93	6.60	1.80	2.08	0.31	2.10	0.30	2.10
				PKP	1175.58							
				PKP	1176.06							
				PKiKP	1179.06							
23	2018-04-02T13:40:34.840Z	559.00	65.59	P	588.98	6.70	1.90	2.01	0.45	2.06	0.20	2.06

For each event, Table reports its time, depth, angular distance to the hydrophone, generated seismic phases and their travel times (as predicted by the ray-tracing code), horizontal, depth and origin time uncertainties, estimated uncertainty σ_{prop} in the travel time, estimated uncertainty σ_{th} in the arrival time, the time-pick quality criterion QC and the uncertainty σ , used in equation (10). For the arrivals satisfying the condition $QC < \sigma_{th}$, σ is equal to σ_{th} . Otherwise, the arrival is rejected (not used in the fit) and its uncertainty σ is omitted.

Table 3: Events detected by the M7 hydrophone (33.45° N, 32.40° W, site depth : 3400 m).

	Event time (UTC)	Depth (km)	Δ (°)	Phase	Travel time (s)	σ_p (km)	σ_d (km)	σ_{or} (s)	σ_{prop} (s)	σ_{th} (s)	QC (s)	σ (s)
1	2016-07-29T21:18:24.740Z	196.00	128.26	PKIKP	1122.90	7.5	1.8	2.80	0.30	2.82	0.60	2.82
				PKiKP	1123.54							
2	2016-08-19T07:32:22.710Z	10.00	88.38	P	772.12	8.4	1.7	1.59	0.48	1.66	0.40	1.66
3	2016-08-29T04:29:57.860Z	10.00	36.03	P	421.37	8.4	1.7	1.76	0.70	1.90	5.00	–
4	2016-11-25T14:24:30.710Z	17.00	80.62	P	731.70	5.9	1.7	1.78	0.42	1.83	0.80	1.83
5	2016-12-25T14:22:27.010Z	38.00	85.39	P	753.33	5.4	1.9	1.59	0.36	1.63	0.05	1.63
6	2017-02-21T14:09:04.320Z	595.98	60.45	P	553.13	9.6	3.0	1.83	0.67	1.95	0.05	1.95
7	2017-04-03T17:40:18.560Z	29.00	78.23	P	716.88	7.5	1.8	2.25	0.47	2.30	0.50	2.30
8	2017-04-24T21:38:30.820Z	28.00	75.94	P	704.15	4.9	1.8	1.48	0.40	1.53	0.50	1.53
9	2017-06-14T07:29:04.390Z	93.00	56.69	P	573.72	2.8	2.2	4.34	0.32	4.35	0.20	4.35
10	2017-07-20T22:31:11.260Z	7.00	48.31	P	521.95	4.3	1.7	1.59	0.42	1.64	0.05	1.64
11	2017-09-08T04:49:19.180Z	47.39	58.21	P	589.27	6.2	3.7	2.47	0.58	2.54	1.50	2.54
12	2017-09-19T18:14:38.090Z	48.00	60.33	P	603.86	4.5	1.8	2.32	0.37	2.35	0.10	2.35
13	2017-11-12T18:18:17.180Z	19.00	63.18	P	626.70	4.9	1.7	2.62	0.41	2.65	0.40	2.65
14	2018-01-10T02:51:33.290Z	19.00	48.35	P	520.31	5.8	1.7	1.65	0.49	1.72	3.00	–
15	2018-01-14T09:18:45.540Z	39.00	63.44	P	625.71	7.0	1.8	2.43	0.47	2.48	0.05	2.48
16	2018-01-23T09:31:40.890Z	14.06	75.97	P	706.44	6.0	2.8	1.57	0.58	1.67	0.05	1.67
17	2018-02-16T23:39:39.280Z	22.00	60.95	P	611.29	4.3	1.8	2.49	0.39	2.52	0.50	2.52
18	2018-04-02T13:40:34.840Z	559.00	61.22	P	561.06	6.7	1.9	2.01	0.47	2.06	0.15	2.06

For each event, Table reports its time, depth, angular distance to the hydrophone, generated seismic phases and their travel times (as predicted by the ray-tracing code), horizontal, depth and origin time uncertainties, estimated uncertainty σ_{prop} in the travel time, estimated uncertainty σ_{th} in the arrival time, the time-pick quality criterion QC and the uncertainty σ , used in equation (10). For the arrivals satisfying the condition $QC < \sigma_{th}$, σ is equal to σ_{th} . Otherwise, the arrival is rejected (not used in the fit) and its uncertainty σ is omitted.

Table 4: Results of the fits.

AUH	fit id	γ'_e	γ'_e 95% CI interval	a (s)	χ^2_ν
M2	M2-1	0.083	[0.048 0.118]	0.57	0.41
	M2-2	0.059	[0.021 0.096]	0.33	0.31
M7	M7-1	0.467	[0.382 0.551]	0.52	1.49
	M7-2	0.434	[0.368 0.501]	-0.93	0.54

For each fit, Table reports the estimated drift rate γ'_e , its 95% confidence interval (CI), estimated value of the time offset a and the goodness-of-fit χ^2_ν .

Table 5: Estimation of the drift-rate spread.

AUH	N	n	C	γ'_{mean}	95% interval	σ_n/γ'_{mean} (%)
M2	15	12	455	0.060	[0.045 0.074]	12.08
		9	5005	0.060	[0.035 0.087]	22.12
M7	13	10	286	0.435	[0.402 0.466]	3.86
		8	1287	0.435	[0.392 0.490]	6.04

For each hydrophone, subsets containing 80% and 60% of the recorded arrivals were considered. Table reports the number of available arrivals N , the number of arrivals n composing a subset and the number of fitted subsets C given by the number of combinations formed by choosing (irrespective of the order within a given combination) n items from N items. As explained in the text, each fit yields a different γ'_{red} . From the resulting collection of γ'_{red} (Figure 10), a mean value γ'_{mean} , an interval occupied by 95% of γ'_{red} values and a ratio of the standard deviation σ_n to γ'_{mean} are estimated in each case and reported in Table.

641 **LIST OF FIGURE CAPTIONS**

642 1 (color online) Bathymetric maps showing the deployment areas of (a) the M2 and
 643 (b) the M7 hydrophones. Bathymetric data are extracted from the ETOPO1 global
 644 relief model (Amante and Eakins 2009). Isobaths are separated by 100 m.

645 2 Examples of *P*-waves detected by the M7 hydrophone with (a) high and (b) low
 646 quality criterion (*QC*) (see Table 3). Top row panels show the entire acoustic signals
 647 while the bottom row panels present an arbitrary zoom on the *P*-wave onset. The time
 648 axes correspond to the AUH time. The time origins were shifted with respect to the
 649 synchronization instant to make the reading of time axes easier. (a)-(b) Vertical lines
 650 indicate the predicted arrival times t_{pred} . Although measured in the GPS time, t_{pred}
 651 are indicated to make the presence of the clock drift visually evident. In its absence,
 652 t_{pred} would coincide with (or be very close to) the observed onsets. (c)-(d) Vertical
 653 bars indicate the user time-picks while the width of the gray vertical bands shows the
 654 assigned *QC* value. The sampling frequency is 240 Hz.

655 3 Histograms illustrating the spread in (a) depths, (b) latitudes, (c) longitudes
 656 and (d) angular distances of all generated hypocenters used to estimate the uncer-
 657 tainty in the arrival time of the *P* phase produced by the September 8, 2017 event at
 658 04:49:19.18 UTC. According to the NEIC catalog, the event depth d_0 was 47.39 km,
 659 the latitude θ_0 and the longitude ϕ_0 of its epicenter were 15.02° N and 93.90° W, re-
 660 spectively, corresponding to an angular distance Δ_0 of 57.35° to the M2 hydrophone
 661 location.

662 4 (color online) Generated seismic velocity models compared to the principal model
 663 ak135. (a) The ak135 model. (b)–(d) The three panels compare the ak135 model (thicker
 664 line with filled circles) and the generated models (thinner lines with empty circles) by
 665 zooming at specific depths, indicated in (a) by horizontal dashed lines.

5 *Left panels* : Histograms of travel times for the P phase propagating between
666 the hypocenter of the September 8, 2017 event and the M2 hydrophone. The travel
667 times are calculated under different conditions : (a) using the ak135 model and the
668 hypocenters with depths and angular distances shown in Figure 3; (c) using the velocity
669 models shown in Figure 4 and the hypocentre position listed in the NEIC catalog ($d_0 =$
670 47.39 km and $\Delta_0 = 57.35^\circ$); (e) using all possible pairs “hypocenter/velocity model”
671 formed from the collections of hypocenters and velocity models employed in the first
672 two cases. Vertical dotted lines indicate the travel time predicted by the ak135 model for
673 the hypocenter at (d_0, Δ_0) . Solid line in (e) is the normal probability density function
674 found by fitting the data. *Right panels* : Histograms of the standard deviations σ_{prop}
675 obtained (see text) from the ensemble of the travel times shown in the corresponding
676 left panel. All histograms are normalized to have an area of 1.

6 Global map of the teleseismic events (circles, Table 1) recorded by the M2 (black
679 star) and M7 (white star) hydrophones. Grey-filled circles indicate events detected by
680 both hydrophones while white-filled and black-filled circles indicate events detected
681 by only one of the hydrophones (M2 or M7, respectively). Dashed lines are epicentral
682 distances in degrees measured from the center of the HYDROMOMAR array.

7 Estimation of γ' for the M7 hydrophone. (a) Predicted t_{pred} versus observed t_{obs}
684 arrival times. The dashed line indicates the case of no clock-drift (i.e. $t_{pred} = t_{obs}$). The
685 uncertainties in t_{pred} (Table 3) are not indicated as they are not visible at the scale of
686 the experiment. (b) Observed time differences ($t_{obs} - t_{pred}$). Filled circles indicate events
687 labeled as oceanic in Table 1. (c) Residuals ($t_{obs}/\gamma_e - t_{pred}$), obtained in the fit M7-2,
688 compared with the corresponding uncertainties σ (Table 3) represented by bars (each
689 bar takes an interval from $-\sigma$ to σ). The absolute value of most residuals is smaller
690 than the corresponding σ , which is reflected by the small value of χ_ν^2 . Arrival numbering
691 corresponds to that of Table 3. Oceanic events shallower than 35 km and events not
692 passing the quality criterion are excluded from the fit.

8 Estimation of γ' for the M2 hydrophone. (a) Predicted t_{pred} versus observed t_{obs}
 693 arrival times. The dashed line indicates the case of no clock-drift (i.e. $t_{pred} = t_{obs}$). The
 694 uncertainties in t_{pred} (Table 2) are not indicated as they are not visible at the scale of
 695 the experiment. (b) Observed time differences ($t_{obs} - t_{pred}$). Filled circles indicate events
 696 labeled as oceanic in Table 1. (c) Residuals ($t_{obs}/\gamma_e - t_{pred}$), obtained in the fit M2-2,
 697 are compared with the corresponding uncertainties σ (Table 2) represented by bars
 698 (each bar takes an interval from $-\sigma$ to σ). The absolute value of almost all residuals is
 699 smaller than the corresponding σ , which is reflected by the small value of χ^2_ν . Arrival
 700 numbering corresponds to that of Table 2. Oceanic events shallower than 35 km and
 701 events not passing the quality criterion are excluded from the fit.
 702

9 Comparison of the observed time differences ($t_{obs} - t_{pred}$) with those expected from
 703 equation (1) at moments t_{pred} in case of measured γ'_m (dashed line) and estimated γ'_e
 704 (solid line) drift rates for the events used in (a) fit M2-2 and (b) fit M7-2. The error
 705 bars represent uncertainties σ on t_{pred} and the gray zone is bordered by limiting values
 706 of the 95% confidence interval on γ'_e (as reported in Table 4).
 707

10 Histograms of the γ'_{red} obtained from the fits based on subsets of the arrivals used
 708 in fits M2-2 and M7-2. Each subset contains (a)-(b) 80% of total number of arrivals (12
 709 and 8 in case of the M2 and M7 hydrophones, respectively) and (c)-(d) 60% (11 and 7
 710 in case of the M2 and M7 hydrophones, respectively). In each panel, the vertical line
 711 indicates the value of the measured γ'_m while the darker gray color outlines an interval
 712 occupied by 95% of γ'_{red} values. All histograms are normalized to have an area of 1.
 713 The horizontal axes are identical column-wise.
 714

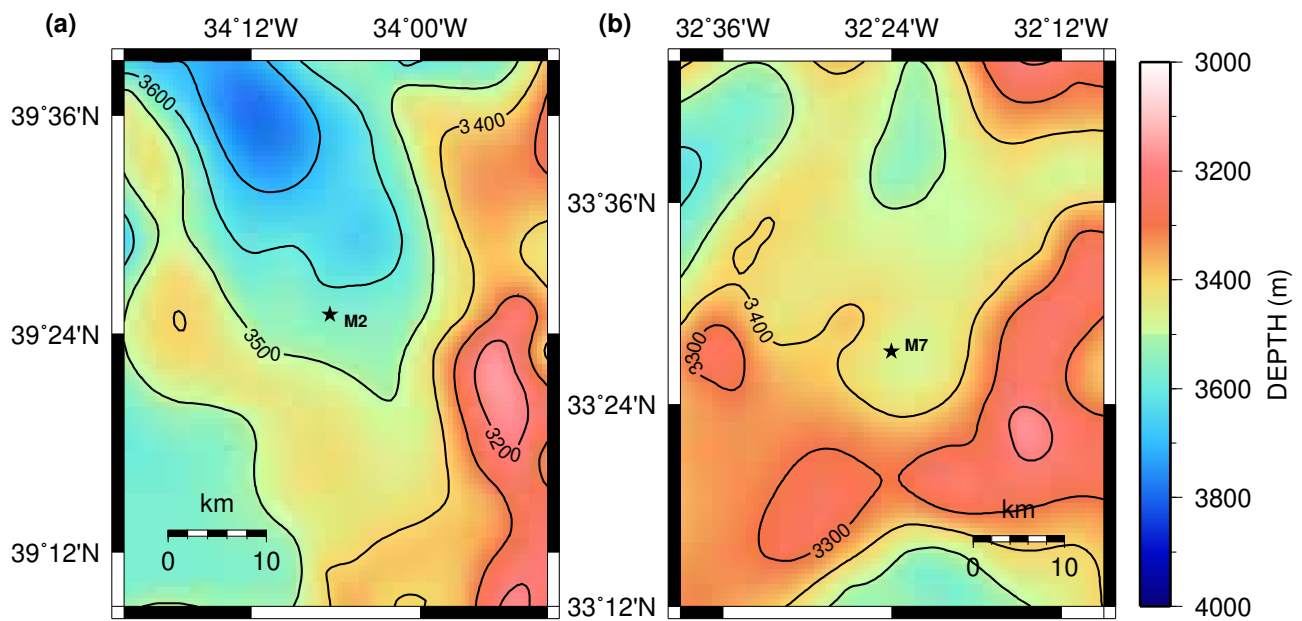


Figure 1. (color online) Bathymetric maps showing the deployment areas of (a) the M2 and (b) the M7 hydrophones. Bathymetric data are extracted from the ETOPO1 global relief model (Amante and Eakins 2009). Isobaths are separated by 100 m.

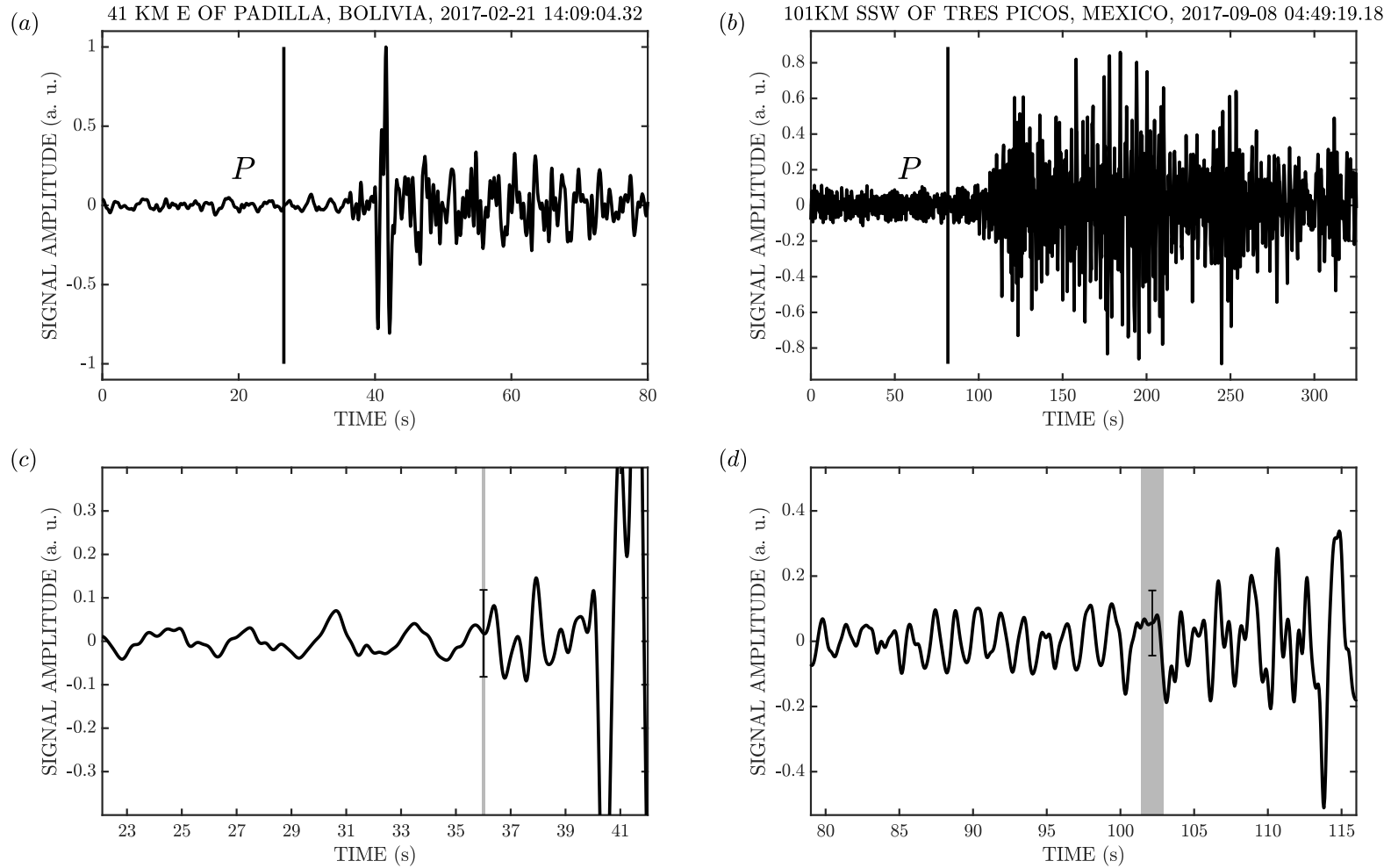


Figure 2. Examples of P -waves detected by the M7 hydrophone with (a) high and (b) low quality criterion (QC) (see Table 3). Top row panels show the entire acoustic signals while the bottom row panels present an arbitrary zoom on the P -wave onset. The time axes correspond to the AUH time. The time origins were shifted with respect to the synchronization instant to make the reading of time axes easier. (a)-(b) Vertical lines indicate the predicted arrival times t_{pred} . Although measured in the GPS time, t_{pred} are indicated to make the presence of the clock drift visually evident. In its absence, t_{pred} would coincide with (or be very close to) the observed onsets. (c)-(d) Vertical bars indicate the user time-picks while the width of the gray vertical bands shows the assigned QC value. The sampling frequency is 240 Hz.

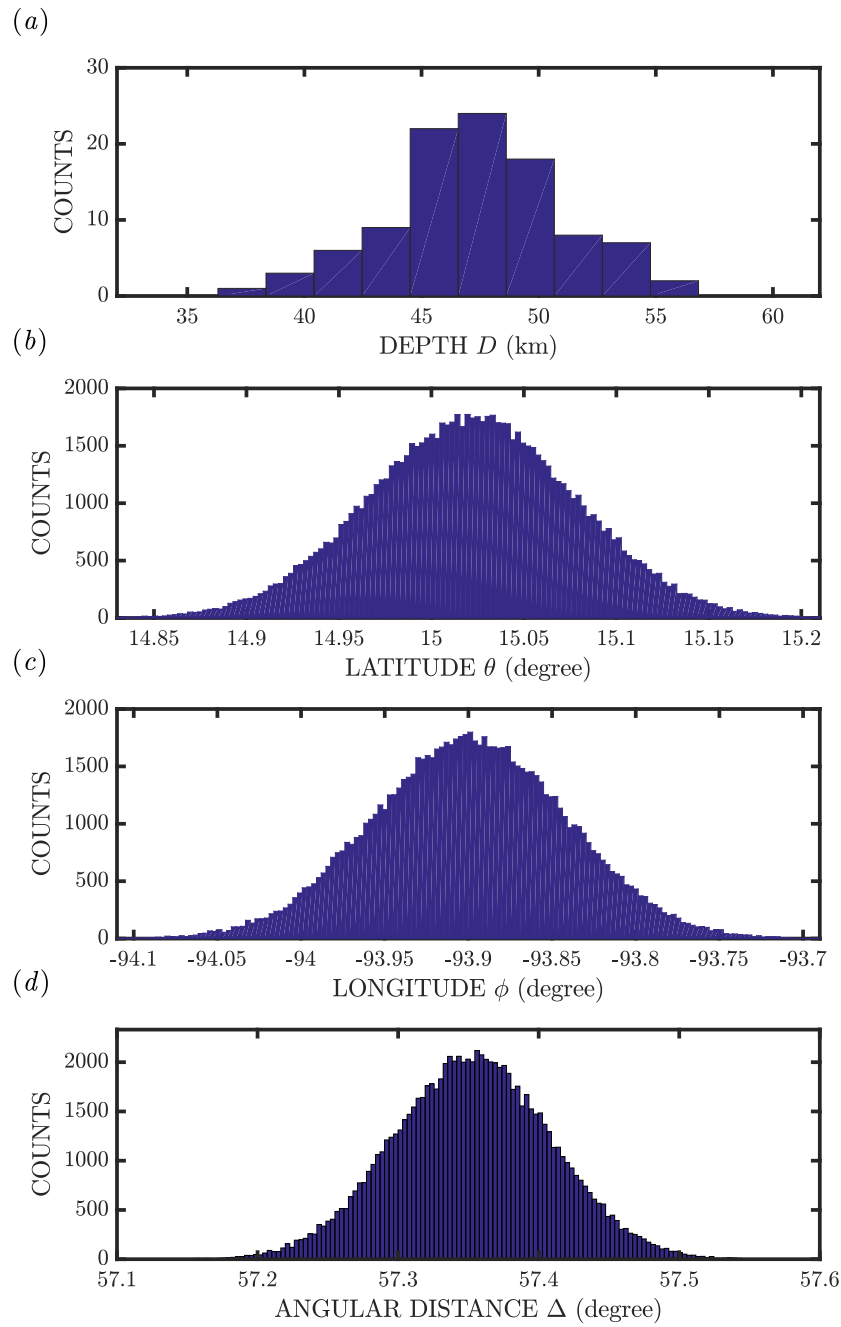


Figure 3. Histograms illustrating the spread in (a) depths, (b) latitudes, (c) longitudes and (d) angular distances of all generated hypocenters used to estimate the uncertainty in the arrival time of the P phase produced by the September 8, 2017 event at 04:49:19.18 UTC. According to the NEIC catalog, the event depth d_0 was 47.39 km, the latitude θ_0 and the longitude ϕ_0 of its epicenter were 15.02° N and 93.90° W, respectively, corresponding to an angular distance Δ_0 of 57.35° to the M2 hydrophone location.

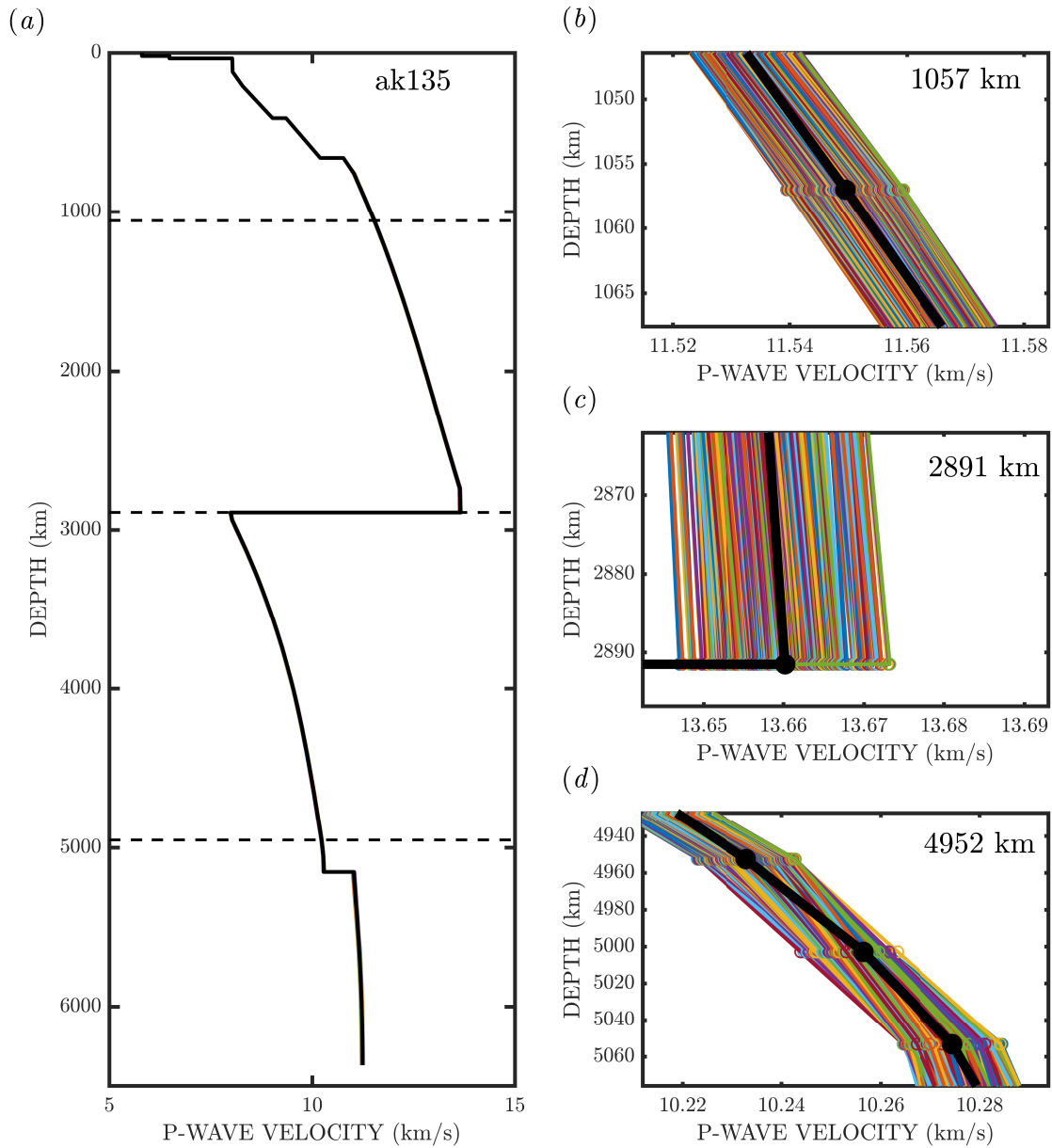


Figure 4. (color online) Generated seismic velocity models compared to the principal model ak135. (a) The ak135 model. (b)–(d) The three panels compare the ak135 model (thicker line with filled circles) and the generated models (thinner lines with empty circles) by zooming at specific depths, indicated in (a) by horizontal dashed lines.

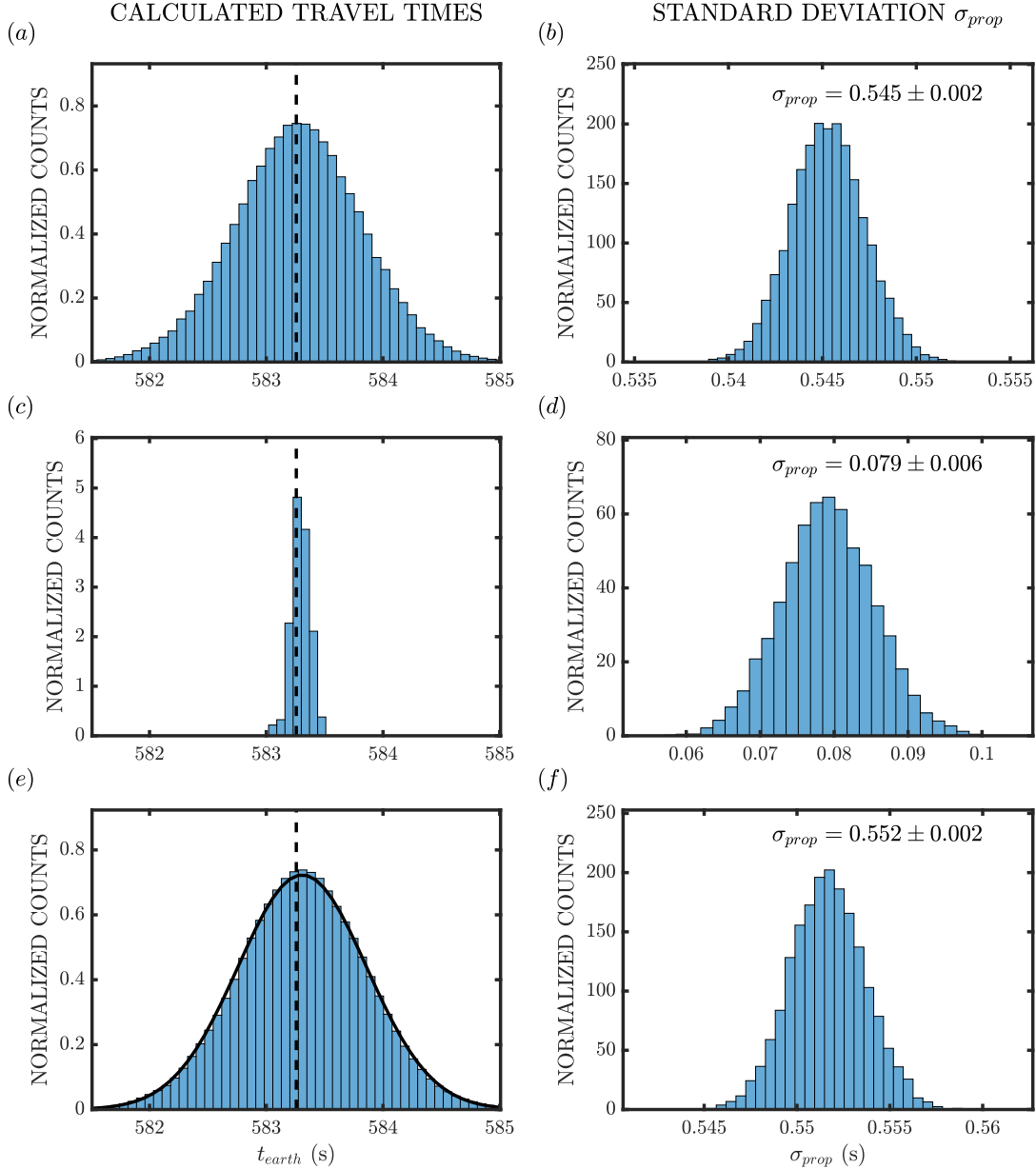


Figure 5. *Left panels* : Histograms of travel times for the P phase propagating between the hypocenter of the September 8, 2017 event and the M2 hydrophone. The travel times are calculated under different conditions : (a) using the ak135 model and the hypocenters with depths and angular distances shown in Figure 3; (c) using the velocity models shown in Figure 4 and the hypocentre position listed in the NEIC catalog ($d_0 = 47.39$ km and $\Delta_0 = 57.35^\circ$); (e) using all possible pairs “hypocenter/velocity model” formed from the collections of hypocenters and velocity models employed in the first two cases. Vertical dotted lines indicate the travel time predicted by the ak135 model for the hypocenter at (d_0, Δ_0) . Solid line in (e) is the normal probability density function found by fitting the data. *Right panels* : Histograms of the standard deviations σ_{prop} obtained (see text) from the ensemble of the travel times shown in the corresponding left panel. All histograms are normalized to have an area of 1.

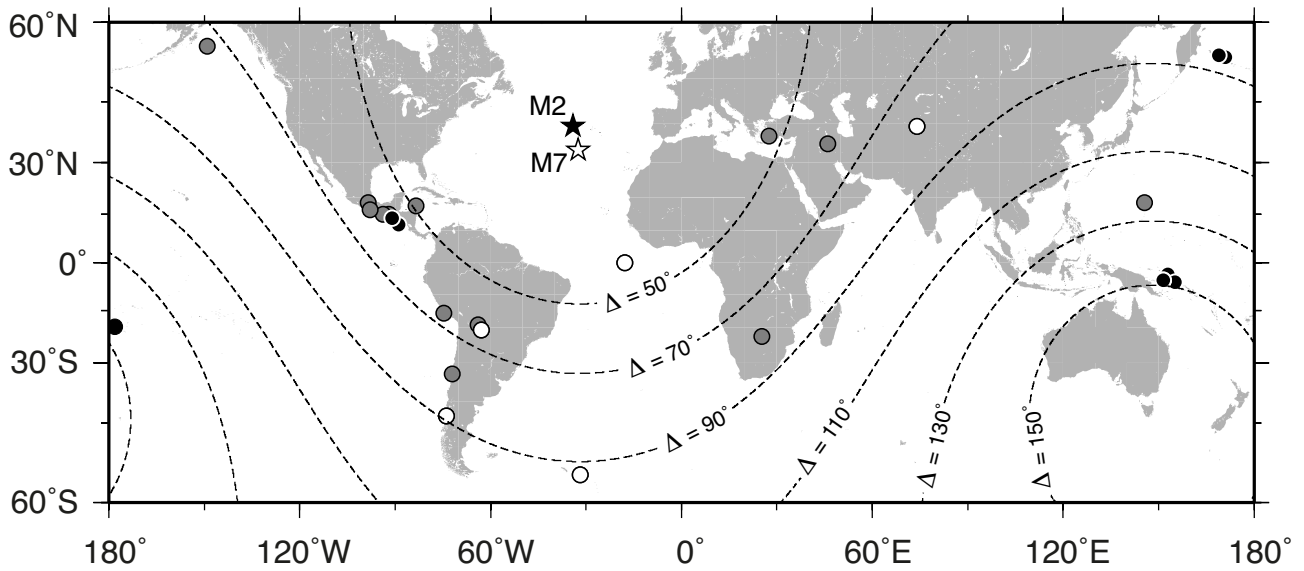


Figure 6. Global map of the teleseismic events (circles, Table 1) recorded by the M2 (black star) and M7 (white star) hydrophones. Grey-filled circles indicate events detected by both hydrophones while white-filled and black-filled circles indicate events detected by only one of the hydrophones (M2 or M7, respectively). Dashed lines are epicentral distances in degrees measured from the center of the HYDROMOMAR array.

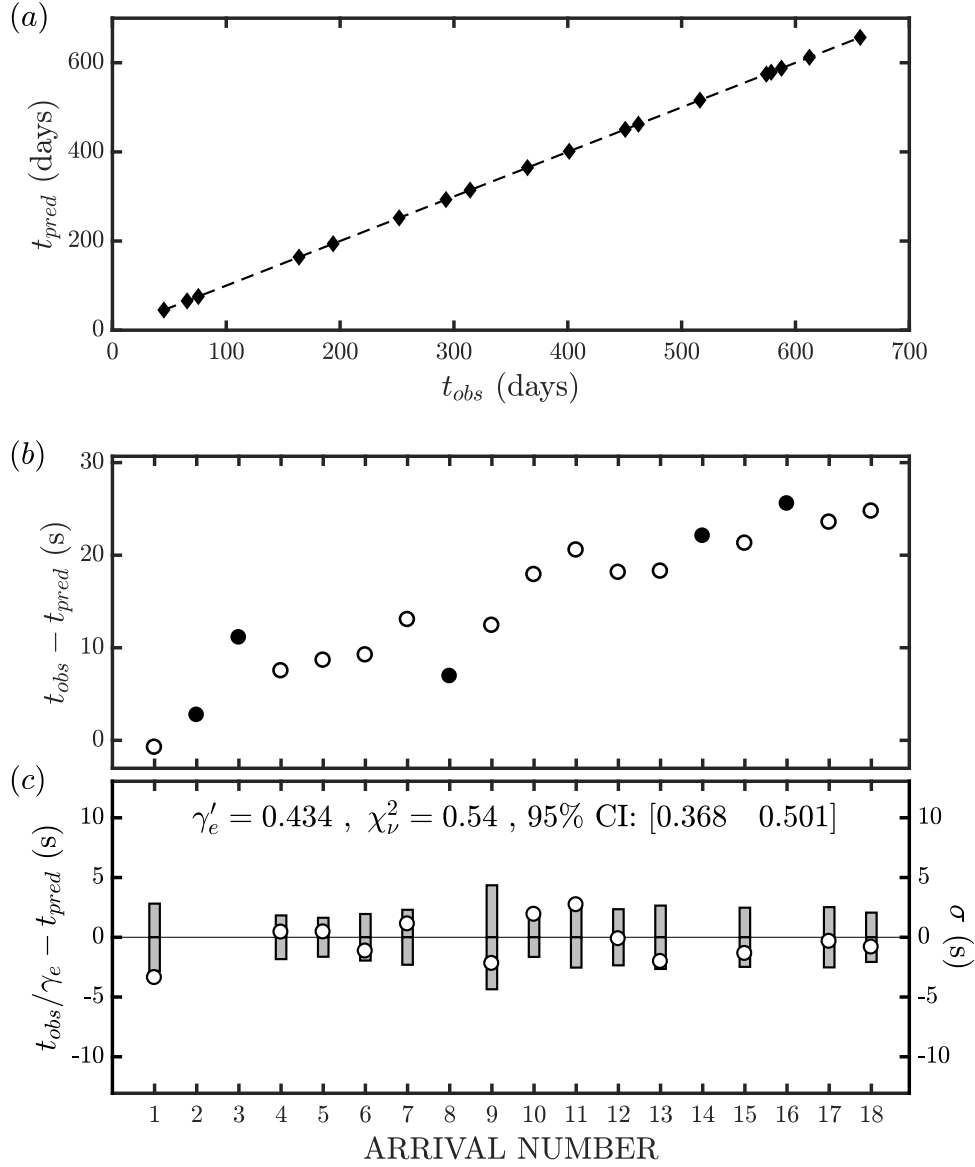


Figure 7. Estimation of γ' for the M7 hydrophone. (a) Predicted t_{pred} versus observed t_{obs} arrival times. The dashed line indicates the case of no clock-drift (i.e. $t_{pred} = t_{obs}$). The uncertainties in t_{pred} (Table 3) are not indicated as they are not visible at the scale of the experiment. (b) Observed time differences ($t_{obs} - t_{pred}$). Filled circles indicate events labeled as oceanic in Table 1. (c) Residuals ($t_{obs}/\gamma_e - t_{pred}$), obtained in the fit M7-2, compared with the corresponding uncertainties σ (Table 3) represented by bars (each bar takes an interval from $-\sigma$ to σ). The absolute value of most residuals is smaller than the corresponding σ , which is reflected by the small value of χ^2_ν . Arrival numbering corresponds to that of Table 3. Oceanic events shallower than 35 km and events not passing the quality criterion are excluded from the fit.

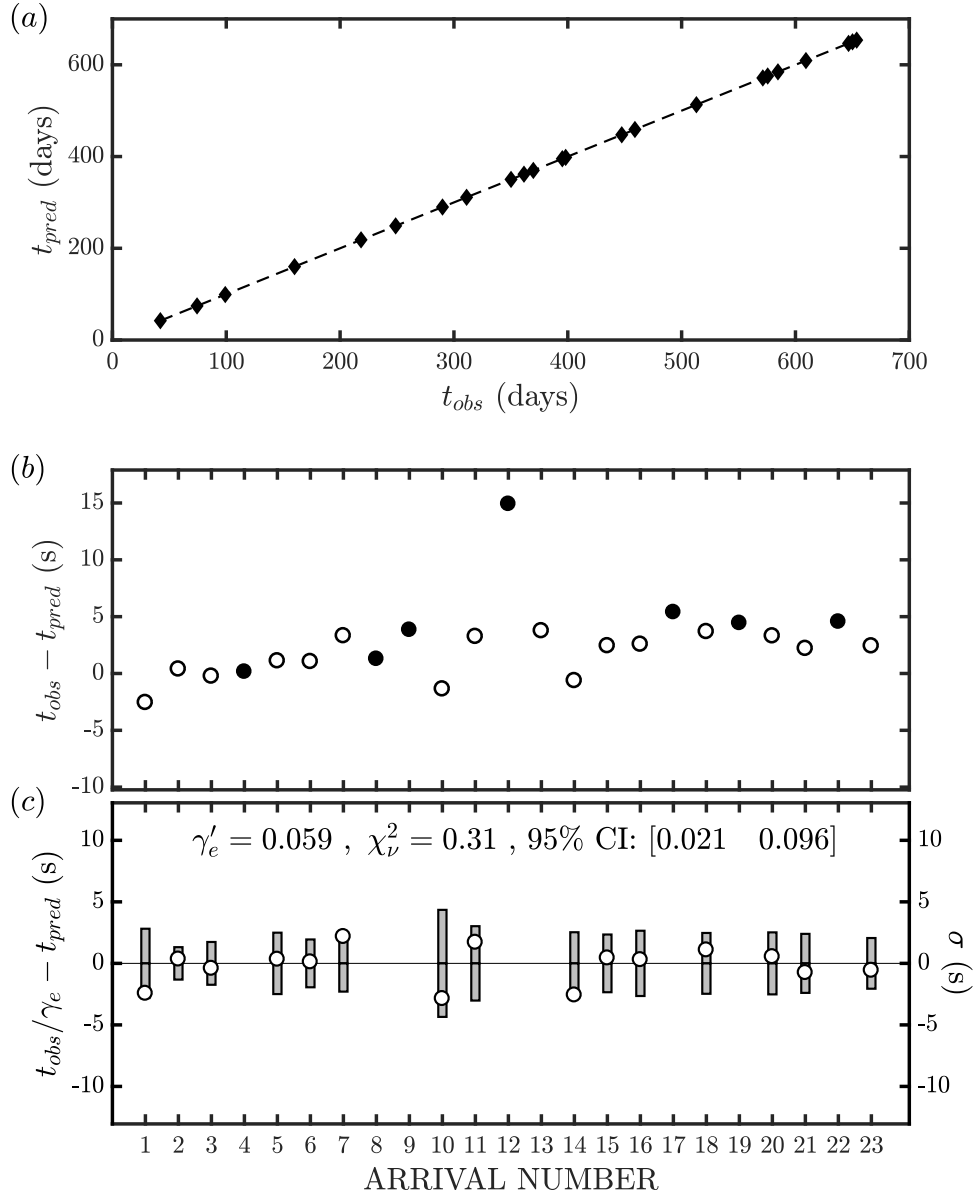


Figure 8. Estimation of γ' for the M2 hydrophone. (a) Predicted t_{pred} versus observed t_{obs} arrival times. The dashed line indicates the case of no clock-drift (i.e. $t_{pred} = t_{obs}$). The uncertainties in t_{pred} (Table 2) are not indicated as they are not visible at the scale of the experiment. (b) Observed time differences ($t_{obs} - t_{pred}$). Filled circles indicate events labeled as oceanic in Table 1. (c) Residuals ($t_{obs}/\gamma_e - t_{pred}$), obtained in the fit M2-2, are compared with the corresponding uncertainties σ (Table 2) represented by bars (each bar takes an interval from $-\sigma$ to σ). The absolute value of almost all residuals is smaller than the corresponding σ , which is reflected by the small value of χ^2_ν . Arrival numbering corresponds to that of Table 2. Oceanic events shallower than 35 km and events not passing the quality criterion are excluded from the fit.

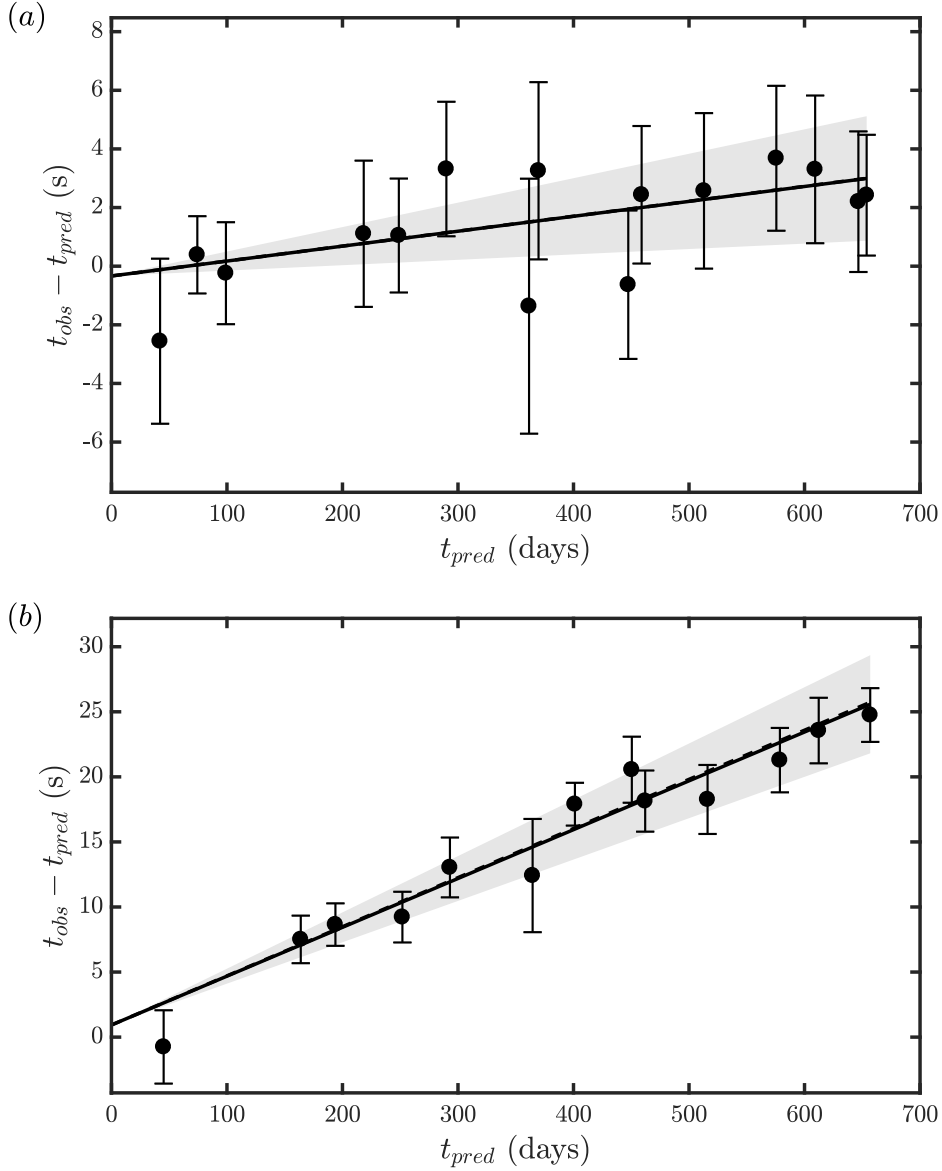


Figure 9. Comparison of the observed time differences ($t_{obs} - t_{pred}$) with those expected from equation (1) at moments t_{pred} in case of measured γ'_m (dashed line) and estimated γ'_e (solid line) drift rates for the events used in (a) fit M2-2 and (b) fit M7-2. The error bars represent uncertainties σ on t_{pred} and the gray zone is bordered by limiting values of the 95% confidence interval on γ'_e (as reported in Table 4).

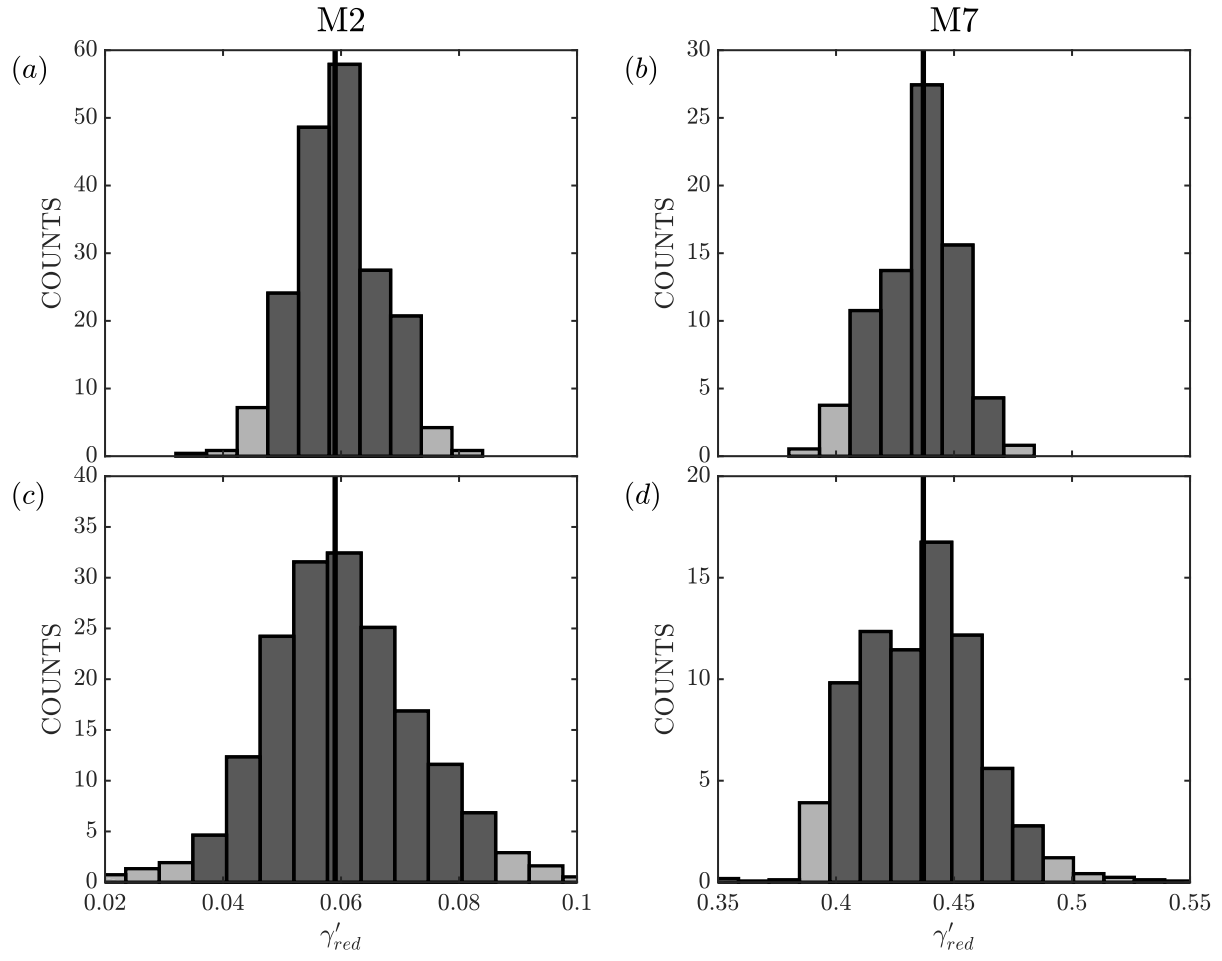


Figure 10. Histograms of the γ'_{red} obtained from the fits based on subsets of the arrivals used in fits M2-2 and M7-2. Each subset contains (a)-(b) 80% of total number of arrivals (12 and 8 in case of the M2 and M7 hydrophones, respectively) and (c)-(d) 60% (11 and 7 in case of the M2 and M7 hydrophones, respectively). In each panel, the vertical line indicates the value of the measured γ'_m while the darker gray color outlines an interval occupied by 95% of γ'_{red} values. All histograms are normalized to have an area of 1. The horizontal axes are identical column-wise.