Seismic Velocity Response to Atmospheric Pressure Using Time-Lapse Passive Seismic Interferometry

Chloé Gradon, Florent Brenguier, Johannes Stammeijer, Aurélien Mordret,
Kees Hindriks, Xander Campman, Richard Lynch, Pierre Boué, Malgorzata
Chmiel

► To cite this version:

Chloé Gradon, Florent Brenguier, Johannes Stammeijer, Aurélien Mordret, Kees Hindriks, et al..
Seismic Velocity Response to Atmospheric Pressure Using Time-Lapse Passive Seismic Interferometry.
Bulletin of the Seismological Society of America, 2021, 10.1785/0120210069 . hal-03320516

HAL Id: hal-03320516

<https://hal.science/hal-03320516>

Submitted on 16 Aug 2021

HAL is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.

1 Seismic velocity response to atmospheric pressure
2 using time-lapse passive seismic interferometry

3 Chloé Gradon (UGA, Sisprobe), Florent Brenguier (UGA),
Johannes Stammeijer (PDO), Aurélien Mordret (UGA),
Kees Hindriks (Shell), Xander Campman (Shell),
Richard Lynch (Sisprobe), Pierre Boué (UGA)
and Malgorzata Chmiel (ETH)

4 August 12, 2021

5 Corresponding Author: Chloé Gradon (chloe.gradon@univ-grenoble-alpes.fr),
6 Université Grenoble Alpes, ISTerre, CS 40700, 38058 GRENOBLE Cedex 9,
7 France

8

9

10

11 **Declaration on competing interest**

12 The authors acknowledge there are no conflicts of interest recorded.

13

14

15 **Abstract**

16 Seismic velocities in the shallow crust down to a few kilometers depth
17 show a remarkable sensitivity to stress perturbations due to the presence
18 of compliant pores, cracks, fractures, and faults. Monitoring temporal

changes of seismic velocities can thus provide key insights on dynamic processes affecting the shallow crust such as those related to the atmosphere (rainfall, barometric pressure, temperature) and those with deeper tectonic and volcanic origins. In this work we investigate the specific response of the near-surface down to 300m depth to atmospheric pressure variations. We conduct a four-month passive seismic monitoring experiment in the desert of Oman using continuous noise recorded at geophones located within five wells. The results show a clear, direct correlation between seismic velocities and barometric pressure variations for monthly transients. At a longer, seasonal temporal scale, seismic velocities are stable while atmospheric pressure shows a clear positive trend. We use the undrained coupled poro-elastic theory to model these observations and find that the lack of seasonal velocity changes can be partly explained by the atmospheric pressure that diffuses into the pores with a strong hydraulic diffusivity likely higher than $100m^2/s$ consistent with the local geology referring to carbonates. Finally, the comparison between the modeled and observed velocity changes leads to estimate a velocity-stress sensitivity on the order of $6.3 \cdot 10^{-7} Pa^{-1}$ which is consistent with previous studies. Using this result for calibration, we find that a sudden step-change drop of velocity of 0.015% occurring in the beginning of October and corresponding to a stress perturbation likely larger than $240Pa$ affected the entire studied area. This small change could be related to a perturbation at greater depth associated with variations in the production rates within the underlying reservoir.

1 Introduction

Improving our ability to monitor stress changes occurring at seismogenic depth is crucial to better understand processes associated with the generation of large earthquakes (Kato & Ben-Zion, 2021). Both laboratory and field experiments have shown that stress perturbations as small as a few kiloPascal have a measur-

48 able impact on seismic velocities (e.g. Yamamura et al. (2003)). This stress de-
 49 pendence is known to be mostly controlled by the opening/closing of microcracks
 50 due to changes in effective pressure. Continuous seismic velocity monitoring has
 51 been successfully applied to study controlled fluid injection, environmental, tec-
 52 tonic and volcanic processes both at a laboratory scale (Scuderi & Collettini,
 53 2016), and in nature using controlled sources (Niu et al. (2008), Tsuji et al.
 54 (2018), Doetsch et al. (2018)) and ambient seismic noise (Sens-Schönfelder &
 55 Wegler (2006), Brenguier et al. (2008a)).

56 Better understanding velocity changes induced by environmental perturba-
 57 tions is key in unveiling transients associated with deeper tectonic or volcanic
 58 processes. These environmental, exogenous processes show a variety of time
 59 scales ranging from hours for tidal-induced strain (Yamamura et al., 2003) to
 60 days or months for rainfall-induced pore pressure changes (Rivet et al. (2015),
 61 Clements & Denolle (2018)) and atmospheric pressure loading (Silver et al.,
 62 2007) with a strong seasonal, yearly component (Wang et al., 2017). In this
 63 paper we focus on the effects of atmospheric, barometric pressure temporal
 64 changes on near-surface seismic velocities following the seminal paper of Silver
 65 et al. (2007). In order to isolate the atmospheric pressure effect we conduct a
 66 noise-based, passive seismic monitoring experiment (Sens-Schönfelder & Wegler
 67 (2006), Brenguier et al. (2008a)) in an arid location in the middle of the Oman
 68 desert thus strongly reducing the potential effects of rainfall activity or oceanic
 69 tidal loading.

70 We monitor seismic velocities following a workflow similar to Olivier et al.
 71 (2015) and measure temporal changes of the travel time of coda-waves retrieved
 72 from noise-correlations between strings of geophones located in 5 different wells
 73 in the frequency range 10 to 50Hz. By using four months of continuous seismic
 74 records from 56 borehole geophones located down to 365m depth, we find a
 75 consistent trend of seismic velocity change over the entire study area. While
 76 the short-term transients (a few weeks) of velocity changes show a striking

77 similarity with atmospheric pressure variations, we do not observe any long-
 78 term, seasonal trend as seen in the atmospheric pressure data. We use the
 79 model of Roeloffs (1996) to evaluate the effective pressure in the saturated near-
 80 surface using the fully coupled (Biot) equations of elastic deformation and pore
 81 fluid flow taking the atmospheric pressure as the source of aquifer loading. By
 82 showing a long-term equilibrium between the confining and the pore pressure,
 83 this model succeeds in explaining the reduction of seasonal velocity changes
 84 and provides a framework for assessing poroelastic parameters such as hydraulic
 85 diffusivity. Finally, we use this model to calibrate the observations of seismic
 86 velocity variations and find a step-change velocity drop that could be related to
 87 production-rate variations at the deeper reservoir.

88 **2 Data processing**

89 We use a dataset composed of 4 months of continuous seismic records from 56,
 90 three-component geophones spread over five observation wells located above an
 91 oil reservoir in the desert of the Sultanate of Oman (Figure 1, see Data and
 92 resources section). The top geological unit is mostly composed of carbonates
 93 with an average P-wave velocity of 2625 m/s down to about 250 m depth,
 94 followed by alternating carbonate and shale layers and a domed reservoir at
 95 an average depth of about 370m. An aquifer is also present with a water level
 96 recorded at around 50m below the surface during the experiment.

97 The geophones are distributed in five seismic monitoring wells (SMW). Four
 98 wells are located close to the production area (SMW1, SMW2 SMW3, and
 99 SMW4) and one well is about 1km away (SMW5). The minimum and maximum
 100 distances between two wells are 255m and 1775m, respectively (Figure 1-a).
 101 There are eight geophones in each well with an average spacing of 20m, covering
 102 depths ranging from 190 to 330m below the surface (Figure 1-b and c). The
 103 dataset consists of 122 days (four months) of ambient seismic noise recorded

104 from the beginning of August to the end of November 2019. The raw data was
 105 recorded at 4kHz and downsampled at 250Hz to reduce computational cost.

106 Figure 2 shows the averaged daily spectrograms for the Z component of
 107 two sensors in wells SMW3 and SMW5. These spectrograms are obtained by
 108 summing daily spectrograms over the 122 days of the study, highlighting the
 109 average noise variations over a day. In this example, we show data from the
 110 deepest sensors in each well, located at 321 and 362 meters below the surface,
 111 respectively. For both sensors, most of the energy is observed below 50Hz. We
 112 explain the difference of amplitudes between both wells by the fact that SMW3
 113 is located above the production area while SMW5 is farther away. The decrease
 114 of energy during the nights and lunch breaks indicates that the dominant source
 115 of noise is related to human activity around the production area. The emitted
 116 noise is strong enough to be clearly measured at the bottom of the wells above
 117 the production area. We also see clearly monochromatic components at 50 and
 118 100Hz associated with the electrical network.

119 We first cross-correlate the 122 days of data using 15-minute time windows
 120 and a spectral normalization scheme to minimize the impact of the microseis-
 121 micity present in the area. We only compute correlations over the ZZ, EE and
 122 NN components, which are independent, to optimize signal to noise ratio with-
 123 out increasing computational costs. We work between 10Hz and 50Hz to take
 124 advantage of the increased energy in this frequency band. Figure 3-a shows
 125 the cross-correlations (CC) according to calendar time on the y-axis for the EE
 126 component between the shallowest sensor of SMW3 and the shallowest sensor
 127 of SMW4. Each row corresponds to a 15-minute cross-correlation, with col-
 128 ors representing amplitude. The red and black dotted lines are the theoretical
 129 arrival for the P and S-wave in the first layer, respectively. By studying the
 130 variation in the cross-correlations over 3 days, we can observe two significantly
 131 different waveforms that are correlated with active and quiet periods on the pro-
 132 duction site. In order to obtain more stable results, we focused on "nighttime"

133 cross-correlations corresponding to a period ranging from 20h00 to 05h00.

134 We apply the Moving Window CS method (MWCS), also called doublet
135 method, (Poupinet et al. (1984), Brenguier et al. (2008b)) on the negative and
136 positive part of the coda of the correlations to estimate relative travel time
137 variations (dt/t) in the medium. Typically, dt/t time series are obtained by
138 measuring the arrival time difference (dt) between wavepackets in a test trace,
139 i.e. a trace associated to a calendar time T , and a reference trace. In the MWCS
140 method, we measure the dt for a set of time windows in the coda of the traces.

141 The reference trace used to measure velocity changes is usually the stack of
142 all the good-quality cross-correlations over the study period. This reference must
143 be similar enough to each test trace to allow for time shift measurements. In this
144 study, we use the normalized correlation coefficient between the reference trace
145 and the test traces to measure this similarity. We estimate empirically that a
146 correlation coefficient of 0.8 is necessary to obtain reliable results. Studying the
147 evolution of the cross-correlation over the 122 nights (Figure 3-b), we observe
148 that we have remarkable stability over sub-periods ranging from 3 weeks to
149 1.5 month. There is also a very clear change in the cross-correlation waveform
150 between each stable sub-period, allowing us to identify them visually (horizontal
151 dotted lines).

152 Preliminary tests show that using a single reference over the four-month
153 period of the study will not allow us to satisfy the criteria of high coherence
154 between the test and the reference correlations. The clear changes in the corre-
155 lation traces in Figure 3-b, easily explain why that is the case. Reviewing the
156 correlation coefficient for every sensor pairs, we determined a set of subperiods
157 where the reference and test traces showed sufficient similarity. In Figure 3-b,
158 the references associated to each subperiod are plotted in black, blue and red,
159 with their associated correlation coefficient with test traces obtained by stack-
160 ing the cross-correlation over three nights represented in Figure 3-c. We chose
161 three-night stacks as a compromise between stability and time resolution.

162 We apply the MWCS method over each subperiod using non-overlapping
 163 three-night stacks. The coda part of the trace used here starts 0.2s after the
 164 theoretical arrival time of the S-wave and end at a lapse time of two seconds. We
 165 measure the time-shifts in the 10 to 50Hz frequency band, and use 0.4-second
 166 time windows that correspond to four periods of the lower frequency signal. To
 167 simplify our analysis, we compute the dt/t for each sensor pair over components
 168 ZZ, EE and NN (thin lines figure 4) and derive one average per well pair (thick
 169 lines). We finally make the assumption that velocity changes are homogeneous
 170 over the region sampled by these coda-waves (Breguier et al. (2008a)) and use
 171 the standard relation $dt/t = -dv/v$ to convert relative travel time into relative
 172 velocity changes. As the averaged velocity measurements overlap, we are able to
 173 reconstruct the total velocity variations by matching the values at the beginning
 174 and end of each subperiods, which are represented in Figure 4 by different colors.
 175 In practice, these values match without the need for correction in most cases.
 176 The gaps in the measurements are due to missing data or lack of high-quality
 177 correlations.

178 To make sure that using several references does not affect a possible long
 179 term trend in the dv/v measurements, we used a single reference and 17-night
 180 correlations stacks. At this resolution, the similarity between the reference over
 181 the four months and test trace was sufficient to obtain stable results. The trend
 182 using a single or multiple references is the same, showing that no long term bias
 183 was introduced using our method.

184 **3 Results and interpretation**

185 For each well we compute an average velocity variation over all pairs of sen-
 186 sors/components and derive one velocity variation per well by averaging all
 187 its associated velocity change measurements (Breguier et al. (2014), Sens-
 188 Schönfelder et al. (2014)). The results are displayed in Figure 5-b using black

189 lines. The grey area represents the uncertainties on the measurement obtained
 190 by averaging the standard deviations of sensor pair measurements between wells.
 191 By considering studies of the spatial sensitivity of coda-wave travel-time change
 192 measurements (Snieder et al. (2019)) we evaluate that our observations of Fig-
 193 ure 5-b relate to mostly S-wave velocity changes within a region of about 500m
 194 around the wells at an average sensor depth of about 260m. The sensitivity of
 195 each velocity change time series thus overlap spatially and cover most of the
 196 studied area shown on Figure 1. The great similarity of velocity change time se-
 197 ries between all wells of Figure 5-b supports the interpretation of a homogeneous
 198 velocity change over the entire studied area.

199 During the studied period, the global production rates over the site are sta-
 200 ble. No rainfall was reported and we found that stress perturbations from both
 201 tidal oceanic loading and solid earth tides are an order of magnitude smaller
 202 than atmospheric pressure variations after smoothing using a three-day averag-
 203 ing moving window. Figure 5-a shows the in situ barometric pressure. It shows
 204 peak-to-peak diurnal variations of about 500Pa, middle-term changes with peri-
 205 odicity from a few days to a few weeks of maximum 1kPa amplitude and a long-
 206 term, seasonal increase of about 2.5kPa. For comparison with rainfall-induced
 207 loading, 1kPa corresponds to the loading from 100mm of rainfall which is quite
 208 strong meaning that atmospheric pressure loading might be often dominant in
 209 the response of the shallow crust to environmental changes.

210 Interestingly, barometric pressures and seismic velocities show a direct corre-
 211 lation from the beginning of August to mid-September during a strong transient
 212 of barometric pressure change lasting about 30 days (minima around August 10
 213 and September 10 and maximum around August 20). The response of a porous
 214 medium to loading from atmospheric pressure variations has been described by
 215 Van Der Kamp & Gale (1983). Using the constitutive relations for an isotropic
 216 coupled fluid-saturated porous elastic medium from Rice & Cleary (1976), Van
 217 Der Kamp & Gale (1983) showed that under undrained conditions, meaning

without considering fluid flow and pore pressure migration, pore pressure (Po) is directly proportional to the confining pressure (Pc) induced by barometric pressure loading. The coefficient of proportionality, ranging from 0 to 1, is also called the Skempton's coefficient α with a value of 0.5 ($Po = \alpha Pc$). This value is low compared to what is found in laboratory measurements (Ahmadinejad & Kivi, 2021), but is still realistic in an in-situ context (Roeloffs, 1996). Doetsch et al. (2018) observed that seismic velocity changes (dv/v) were proportional to effective pressure $Pe = Pc - Po$, where Po is pore pressure. Utilizing Skempton's coefficient leads to $dv/v \propto Pe = Pc(1 - \alpha)$.

This direct proportionality between velocity changes and the confining, atmospheric pressure can explain the early velocity transient, but long-term velocity changes do not show the increasing trend this theory would predict considering the rise of atmospheric pressure from August to December. The lack of long-term velocity increase suggests a process leading to an increase of pore pressure associated with a decrease of velocity. This situation is typically reached under drained conditions involving pore pressure migration (Roeloffs, 1996). Following the same constitutive laws as Van Der Kamp & Gale (1983), Roeloffs (1996) studies the effect of pore pressure variations induced by atmospheric pressure changes in the case of a vertical, downward migration of fluids from the top of the water table that is directly connected to the atmosphere. In this case, pneumatic diffusion (diffusion of the air in the pores in the unsaturated zone) is neglected (see sketch on Figure 6).

Equation 37 of Roeloffs (1996) describes a harmonic solution for pore pressure according to the frequency of atmospheric pressure changes (ω), the depth from the top of the water table (z) and the hydraulic diffusivity (c):

$$Po(z, \omega, c)/Pc(\omega) = \alpha + (1 - \alpha).exp[-\sqrt{i\omega/c}.z] \quad (1)$$

If α is significantly different from 1, as is the case here ($\alpha = 0.5$), we see different

behaviors at short and long periods (or small and large depths). When the depth is large or for short-term transients (large ω), this solution reduces to the undrained case discussed earlier where the pore pressure is directly proportional to the confining, atmospheric pressure (Figure 6-a). For small depth or long-term pressure variations, the second term representing pore pressure migration dominates and eventually tends to the boundary condition at $z=0$ for which the pore pressure equals the confining pressure thus leading to a 0 net change of seismic velocities (Figure 6-b).

In order to directly compare this model to our observations we derive a synthetic velocity change time series by using the following relation:

$$dv/v = K.(Pc - Po)$$

where the effective pressure $Pc - Po$ is computed as the difference between the atmospheric pressure and the atmospheric pressure-induced pore pressure. K is the sensitivity of velocity changes to effective pressure perturbations in Pa^{-1} . To get Po , we convolve the impulse response of the harmonic solution for pore pressure variation of eq. 1 to the in-situ atmospheric pressure variation time series. We computed this model for a coarse range of depths and diffusivity coefficients. Figure 5-b shows in red the best fit of this model to the observed velocity changes at 50m below the water table level ($z=100m$ below the surface). This is above the first sensors in the wells, and could be due to a higher sensitivity of our method to velocity changes at shallower depth. Indeed, the scattering is expected to be stronger close to the surface, resulting in better sampling of the medium by the coda waves used in our measurements. The associated value for the diffusivity is of the order of $100m^2/s$, which is consistent with values found in the literature for carbonates (Roeloffs, 1996). The estimated value for the sensitivity of velocity changes to effective pressure changes over the first month were the correlation between our model and the measurement is the best is $K = 6.3.10^{-7} Pa^{-1}$. This value is typically in the range of values previously

269 observed in-situ for environmental changes, with sensitivity ranging from 10^{-9}
 270 to 10^{-6} measured by different teams (Yamamura et al., 2003).

271 It is important to note that our model is not well constrained. Figure 7
 272 shows different models associated with 9 combinations of depths ($z = [100, 150,$
 273 $250]\text{m}$) and diffusivities ($c = [10, 100, 1000]\text{m}^2/\text{s}$). While most models still
 274 display strong long-term trends, we see that model 6 is very similar to the one
 275 showed Figure 5, and that model 3 shows a better fit. We discarded those models
 276 due to the high value of $c = 1000\text{m}^2/\text{s}$. However, according to Roeloffs (1996),
 277 this number is typical for karsts involving highly diffusive vertical fractures, and
 278 further study would be necessary to better constrain the depth and diffusivity
 279 in this analysis.

280 Interestingly, around October 10, the velocity change observations deviate
 281 from the modeled effective pressure by showing an additional step change drop
 282 of velocity of about $-1.5 \cdot 10^{-4}$. Considering the estimated velocity-pressure
 283 sensitivity discussed above, we find that this velocity drop corresponds to a
 284 reduction of confining pressure or increase of pore pressure of 240Pa. This value
 285 is quite small but the similarity of velocity variations between all wells indicates
 286 that it is likely widespread over the studied area. The wells also overlay an oil
 287 and gas reservoir. It is possible that the changes in production rates are causing
 288 this step-change decrease of velocity by causing a dilatational expansion of the
 289 overburden. We only have access to global rates across the site and cannot
 290 conclude from the available production data. Moreover, such a weak stress
 291 perturbation of 240Pa is probably hardly detectable by other means such as
 292 GPS or InSAR, nor by micro-seismicity monitoring, as it probably does not
 293 affect the integrity of the overburden.

294 4 Conclusions

295 By measuring time-lapse changes of seismic velocities within an aquifer located
296 in an arid desert using passive coda-wave interferometry, we are able to observe
297 velocity changes as small as 0.01% over four months with a temporal resolution
298 of three days. The comparison of our results to barometric pressure time-series
299 shows that the aquifer responds as an undrained poro-elastic medium for short-
300 term barometric pressure variations of a few weeks but that fluid migration
301 plays an important role at a seasonal time scale. The direct correlation between
302 atmospheric loading and velocity variations that we observe is consistent with
303 Silver et al. (2007) active source monitoring experiment in the case of the far-
304 field experiment with wells separated by 30 meters. By allowing the calibration
305 of seismic velocity changes to stress perturbations, we are able to detect a tran-
306 sient effective pressure drop as small as 240Pa over the entire studied area that
307 could be related to changes in the production rates at the underlying reservoir.
308 These results thus provide a framework for monitoring the shallow crust with
309 unprecedented accuracy with applications to active faults or volcano studies.

310 Data and Resources

311 The seismic dataset was provided by Petroleum Development Oman (PDO) and
312 is proprietary. It cannot be released to the public.

313 Acknowledgments

314 We acknowledge PDO and the Game Changer program at Shell for supporting
315 this study, as well as the Ministry of Energy and Minerals of the Sultanate of
316 Oman. This work received funding from the European Research Council (ERC)
317 under Grant Agreement Number 817803 (FaultScan). We are grateful to Michel
318 Campillo and Olivier Coutant for useful discussions, and to the reviewers of this

319 paper for their insights. We also thank Saif Azri for his help.

320 References

- 321 Ahmadinejad, A. & Kivi, I. R., 2021. An experimental investigation on the
322 poroelastic response of a water-saturated limestone to hydrostatic compres-
323 sion, *Bulletin of Engineering Geology and the Environment*, pp. 3817–3832.
- 324 Brenguier, F., Campillo, M., Hadziioannou, C., Shapiro, N., Nadeau, R., &
325 Larose, E., 2008a. Postseismic relaxation along the San Andreas fault at Park-
326 field from continuous seismological observations, *Science*, **321**(5895), 1478–
327 1481.
- 328 Brenguier, F., Shapiro, N. M., Campillo, M., Ferrazzini, V., Duputel, Z.,
329 Coutant, O., & Nercessian, A., 2008b. Towards forecasting volcanic erup-
330 tions using seismic noise, *Nature Geoscience*, **1**(2), 126–130.
- 331 Brenguier, F., Campillo, M., Takeda, T., Aoki, Y., Shapiro, N., Briand, X.,
332 Emoto, K., & Miyake, H., 2014. Mapping pressurized volcanic fluids from
333 induced crustal seismic velocity drops, *Science*, **345**(6192), 80–82.
- 334 Clements, T. & Denolle, M. A., 2018. Tracking Groundwater Levels Using the
335 Ambient Seismic Field, *Geophysical Research Letters*, **45**(13), 6459–6465.
- 336 Doetsch, J., Gischig, V. S., Villiger, L., Krietsch, H., Nejati, M., Amann, F.,
337 Jalali, M., Madonna, C., Maurer, H., Wiemer, S., et al., 2018. Subsurface fluid
338 pressure and rock deformation monitoring using seismic velocity observations,
339 *Geophysical Research Letters*, **45**(19), 10–389.
- 340 Kato, A. & Ben-Zion, Y., 2021. The generation of large earthquakes, *Nature*
341 *Reviews Earth & Environment*, **2**(1), 26–39.

342 Niu, F., Silver, P. G., Daley, T. M., Cheng, X., & Majer, E. L., 2008. Preseis-
343 mic velocity changes observed from active source monitoring at the Parkfield
344 SAFOD drill site, *Nature*, **454**(7201), 204–208.

345 Olivier, G., Brenguier, F., Campillo, M., Roux, P., Shapiro, N., & Lynch, R.,
346 2015. Investigation of coseismic and postseismic processes using in situ mea-
347 surements of seismic velocity variations in an underground mine, *Geophysical*
348 *Research Letters*, **42**(21), 9261–9269.

349 Poupinet, G., Ellsworth, W. L., & Frechet, J., 1984. Monitoring velocity varia-
350 tions in the crust using earthquake doublets: an application to the Calaveras
351 fault, California (USA)., *Journal of Geophysical Research*, **89**(B7), 5719–
352 5731.

353 Rice, J. R. & Cleary, M. P., 1976. Some basic stress diffusion solutions for
354 fluid-saturated elastic porous media with compressible constituents, *Reviews*
355 *of Geophysics*, **14**(2), 227.

356 Rivet, D., Brenguier, F., & Cappa, F., 2015. Improved detection of preeruptive
357 seismic velocity drops at the Piton de La Fournaise volcano, *Geophysical*
358 *Research Letters*, **42**(15), 6332–6339.

359 Roeloffs, E., 1996. Poroelastic Techniques in the Study of Earthquake-Related
360 Hydrologic Phenomena, *Advances in Geophysics*, **38**(C), 135–195.

361 Scuderi, M. M. & Collettini, C., 2016. The role of fluid pressure in induced vs.
362 triggered seismicity: Insights from rock deformation experiments on carbon-
363 ates, *Scientific Reports*, **6**(January), 1–9.

364 Sens-Schönfelder, C. & Wegler, U., 2006. Passive image interferometry and sea-
365 sonal variations of seismic velocities at merapi volcano, indonesia, *Geophysical*
366 *research letters*, **33**(21).

367 Sens-Schönfelder, C., Pomponi, E., & Peltier, A., 2014. Dynamics of Piton de
368 la Fournaise volcano observed by passive image interferometry with multiple
369 references, *Journal of Volcanology and Geothermal Research*, **276**, 32–45.

370 Silver, P. G., Daley, T. M., Niu, F., & Majer, E. L., 2007. Active Source
371 Monitoring of Cross-Well Seismic Travel Time for Stress-Induced Changes,
372 *Bulletin of the Seismological Society of America*, **97**(1B), 281–293.

373 Snieder, R., Duran, A., & Obermann, A., 2019. Locating velocity changes in
374 elastic media with coda wave interferometry, *chap*, **6**, 188–217.

375 Tsuji, S., Yamaoka, K., Ikuta, R., Kunitomo, T., Watanabe, T., Yoshida, Y.,
376 & Katsumata, A., 2018. Secular and coseismic changes in S-wave velocity
377 detected using ACROSS in the Tokai region, *Earth, Planets and Space*, **70**(1).

378 Van Der Kamp, G. & Gale, J. E., 1983. Theory of earth tide and baromet-
379 ric effects in porous formations with compressible grains, *Water Resources*
380 *Research*, **19**(2), 538–544.

381 Wang, Q. Y., Brenguier, F., Campillo, M., Lecointre, A., Takeda, T., & Aoki,
382 Y., 2017. Seasonal Crustal Seismic Velocity Changes Throughout Japan,
383 *Journal of Geophysical Research: Solid Earth*, **122**(10), 7987–8002.

384 Yamamura, K., Sano, O., Utada, H., Takei, Y., Nakao, S., & Fukao, Y., 2003.
385 Long-term observation of in situ seismic velocity and attenuation, *Journal of*
386 *Geophysical Research: Solid Earth (1978–2012)*, **108**(B6).

387 List of Figure Captions

388 **Figure 1:** Seismic Monitoring Well (SMW) locations: a) Top view of the site
389 of the experiments. b) 3D view of the wells. c) Average intersensor distance
390 and depth inside the wells

391

392 **Figure 2:** Averaged daily spectrograms for the deepest sensors of SMW3 (a)
393 and SMW5 (b). Every daily spectrograms is stacked using 15min bins.

394

395 **Figure 3:** Stability analysis of the EE component of the cross-correlations
396 between SMW3 and SMW4 (top sensors). a) Cross-correlations according to
397 calendar time. Each row corresponds to a 15-minute cross-correlation, with
398 colors representing amplitude. Arrows on the x-axis represent the theoret-
399 ical arrival time of S-waves for this sensor pair. b) Background: Nighttime
400 cross-correlation according to calendar time. We distinguish three distinct time
401 periods marked by dashed horizontal lines, corresponding to variations in the
402 cross-correlations. The stacks of the CC for each time period are represented
403 in black, blue and red, respectively. These stacks will be the reference for each
404 time period. c) Cross-correlation coefficient between each reference and 3-night
405 stacked correlations.

406

407 **Figure 4:** Velocity variations measured between SMW4 and SMW5. Each
408 thin line corresponds to a sensor pair. The thick lines are the average velocity
409 variations between the wells. Each color corresponds to a different reference
410 used for the velocity measurement.

411

412 **Figure 5:** a) Pressure variation measured at the surface on the experiment
413 site. b) Red line: effective pressure inverted from atmospheric pressure varia-
414 tions for a diffusivity $c = 100m^2/s$ and a depth $z = 100m$. Black lines: velocity

415 variations averaged over each well. Grey area: average errors on velocity mea-
416 surements.

417

418 **Figure 6:** Effective pressure model. The top and bottom panels illustrate
419 the effects of short and long term atmospheric variations on the pore pressure,
420 respectively.

421

422 **Figure 7:** Representation of effective pressure models (blue lines) computed
423 for various combination of depths and diffusivities. The average velocity change
424 over the five wells (orange lines) was added to each plot for comparison. The es-
425 timated value for the sensitivity of velocity changes to effective pressure changes
426 K computed over the first month is annotated on the bottom right corner of
427 each plot.

428 **Figures**

Figure 1: Seismic Monitoring Well (SMW) locations: a) Top view of the site of the experiments. b) 3D view of the wells. c) Average intersensor distances and depth inside the wells

Figure 2: Averaged daily spectrograms for the deepest sensors of SMW3 (a) and SMW5 (b). Every daily spectrogram is stacked using 15min bins.

Figure 3: Stability analysis of the EE component of the cross-correlations between SMW3 and SMW4 (top sensors). a) Cross-correlations according to calendar time. Each row corresponds to a 15-minute cross-correlation, with colors representing amplitude. Arrows on the x-axis represent the theoretical arrival time of S-waves for this sensor pair b) Background: Nighttime cross-correlation according to calendar time. We distinguish three distinct time periods marked by dashed horizontal lines, corresponding to variations in the cross-correlations. The stacks of the CC for each time period are represented in black, blue and red, respectively. These stacks will be the reference for each time period. c) Cross-correlation coefficient between each reference and 3-night stacked correlations.

Figure 4: Velocity variations measured between SMW4 and SMW5. Each thin line corresponds to a sensor pair. The thick lines are the average velocity variations between the wells. Each color corresponds to a different reference used for the velocity measurement.

Figure 5: a) Pressure variation measured at the surface on the experiment site.
b) Red line: effective pressure inverted from atmospheric pressure variations for a diffusivity $c = 100m^2/s$ and a depth $z = 100m$. Black lines: velocity variations averaged over each well. Grey area: average errors on velocity measurements.

a) Short-term variations (weekly -- high ω)

b) Long-term variations (yearly -- small ω)

Figure 6: Effective pressure model. Panels a and b illustrate the effects of short and long term atmospheric variations on the pore pressure, respectively.

Figure 7: Representation of effective pressure models (blue lines) computed for various combination of depths and diffusivities. The average velocity change over the five wells (orange lines) was added to each plot for comparison. The estimated value for the sensitivity of velocity changes to effective pressure changes K computed over the first month is annotated on the bottom right corner of each plot.