Analysis of signals of a borehole strainmeter in the western rift of Corinth, Greece

Research Article

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Abstract:
This paper presents the first analysis of the records of an elliptical 3-component Sacks-Evertson borehole strainmeter. This high-resolution prototype by the Carnegie Institution of Washington, is installed since 2006 in the western rift of Corinth, Greece. We first present the calibration and the correction from external influences, in order to quantify the detection level of the instrument. We show evidence for pore pressure diffusion from the sea, mostly affecting one component. Neglecting this effect, a first order correction reduces the signal by 90% at tidal periods for 2 components and about 70% for the third one. The residual noise vary from 1 nstrain at 1-hour period to 10 nstrain at 1-day period. It allow to detect slow earthquakes lasting 1 day down to magnitude 4 at an hypocentral distance of 8 kilometers. The uncorrected records at periods smaller than semidiurnal does not reveal any slow strain transient with strong amplitude. During the closest seismic swarm to the site in 2011, the analysis of the records reveals strain steps occuring at the arrivaltimes of seismic waves radiated by the local earthquakes, uncorrelated with the amplitudes and mostly related to dynamic pore pressure instabilities.

Keywords:
- high strain deformation zones
- fractures and faults
- tides and planetary waves
- instrumental noise
- external forcing correction

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1. Introduction

For the last twenty years, many instruments devoted to crustal strain measurements have been installed in seismically active regions, in particular in Japan (tiltmeters in the High-Sensitivity Seismograph Network (Hi-net) array (Okada et al. 2004)) or in western USA (GPS, strainmeters and tiltmeters in the Plate Boundary Observatory (PBO)). The growing sensitivity of instruments such as very-broad band seismometers, borehole strainmeters and tiltmeters, and continuous GPS allows a high-resolution continuous strain records of the seismic cycle, including interseismic, transient phenomena at different periods, from seconds to years. For the last decade, their measurements have boosted the research on transients strain, on their properties and their relationship with seismic activity, in particular following the first discovery of slow slip events in the Cascadia Subduction (Rogers and Dragert 2003) and of non-volcanic tremors in Japan (Obara 2002; Obara and Hirose 2006). New scaling laws relating seismic moment to event duration for these instabilities have been developed, relating small, low frequency earthquakes lasting a few seconds to large slow slip events lasting many months (Ide et al. 2007). Most of these anomalies are found in subduction zones, although transform fault contexts also provide evidence for numerous classes of crustal transients (Linde et al. 1996; Nadeau and Dolenc 2005). In contrast, up to now, very few strain transients have been reported in the extensional context of continental rifting (Bernard et al. 2006). This is quite puzzling, because these areas are often characterized by swarm-like seismic activity, which are usually thought of as triggering processes forced by slow creep or/and by pore pressure migration within a fault system; but the lack of relevant data makes these models hypothetical and unconstrained.

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In the attempt to detect and model strain transients in an extensional tectonic context, strainmeters and continuous GPS have been installed in the western rift of Corinth, in Greece, within the framework of the Corinth Rift Laboratory project (Lyon-Caen et al. 2004). In particular, we installed two Sacks–Everston high resolution borehole strainmeters (Sacks et al. 1971): the first in 2002, a Sacks-Everston dilatometer, near the northern coast of the Gulf, on the Trizonia island; and the second in 2007, a 3-component Sacks–Everston strainmeter, near the village of Monasteraki, 10 km to the west (Fig. 1). This high-resolution strain monitoring complements the 5 continuous GPS and the seismicity monitoring of the CRL arrays. In 2002, the Trizonia dilatometer has provided the first record of a strain transient in the rift, lasting 30 minutes, associated with an unusually shallow (less than 3 km) M=3.5 earthquake on the Psathopyrgos fault system (PSA, Fig. 1), which has been interpreted as a slow creep event on the fault (Bernard et al. 2006). This transient was part of a micro seismic activation of the deepest part of this fault, which lasted 7 weeks starting mid-November 2002. The 10 years records from this dilatometer has been studied more extensively by Canitano (2011), and will be the subject of a subsequent paper.

In the present article, we focus on the Monasteraki strainmeter records, first analyzed by Canitano (2011). After the presentation of the seismotectonic context of the western rift of Corinth, we describe the instrument and characterize some of its systematic, spurious noise (steps and drift). We then report the strain signal response to external forcing (atmospheric pressure, oceanic loading), and to earth tide loading, which allows us to correct for these influences. Finally, we present observation of strain induced by dynamic instabilities occurring after P and/or S waves arrivals. The present article is describing the in-situ calibration of a new-designed borehole strainmeter.

2. The seismic and tectonic context of the Western rift of Corinth

The rift of Corinth is an asymmetric graben, possibly the fastest continental rift on Earth (Armijo et al. 1996; Bernard et al. 2006), with measured geodetically extension varying from 5 to 15 mm/yr between the eastern and western ends (Avalleone et al. 2004; Briole et al. 2000). The Corinth gulf is bounded by large, active E-W striking normal faults mostly outcropping near its southern coast, and dipping to the north (Armijo et al. 1996; Micarelli et al. 2003). In the western part of the rift, which is the focus of the present study, a set of recent en echelon normal faults have connected the Psathopyrgos normal fault which marks the western end of the rift, to the Eastern Helike Fault (Palyvos et al. 2010, 2007). All these north dipping faults root near 5 to 7 km in depth within a highly microseismic layer (Lyon-Caen et al. 2004). The latter dips gently towards the north, and defines a WNW-SEE active trend under the gulf. Some south-dipping, antithetic faults outcrop near the northern coast line, but most do not show any microseismic activity at shallow depth. The geometry of the microseismic structures is now accurately defined thanks to the double-difference relocations of events recorded by the continuous monitoring of the CRL-net seismic array, since 2000 (Lambotte et al. 2010).

The most recent significant earthquakes in the western rift are the Aigion (M=6.2, 1995) (Bernard et al. 1997), and the two 2010 Pyrgos earthquakes (both M=5.3) (Lyon-Caen et al. 2010). The former has activated a low dip angle, north-dipping, blind normal fault. The latter earthquake pair, around 6 to 8 km in depth, coincides with the northern edge of the micro-seismic layer, and may have activated the root of the Psathopyrgos fault (Lyon-Caen et al. 2010). This fault has not ruptured historically (thus for at least 3 centuries) and presently has the potential for a magnitude 6.2, possibly up to 6.7 in case of a dynamic cascade linking neighbouring fault segments (Bernard et al. 2010). Hence, the fault system in the western rift of Corinth may be close to failure at the time scale of a few decades, which may explain its accelerated strain rate (reaching $10^{-5}$/year) and large seismicity rate (more than 10,000 events per year).

The frequent and large seismic swarms of the western rift contrast with the steady strain rate deduced from continuous GPS, suggesting small, or even no aseismic deformation related to this seismicity, at the local resolution of GPS (equivalent moment magnitude $M=5$). However, owing to some diffusion characteristics of space-time evolution of the micro seismicity, it has been proposed that these swarm events may be assisted or triggered by pore pressure diffusion in the fractured, active layer (Bourouis and Bernard 2007), which remains to be quantified.

Figure 1. Map of the western part of the gulf of Corinth (black square, left bottom insert) in Greece. Red diamonds: sites of borehole strainmeter (MOK) and dilatometer (TRIZ). Blue circles: epicenters of 2010 Pyrgos earthquakes, M=5.3, ‘18’: 18/01/2010 event, ‘22’: 22/01/2010 event. Blue diamonds: epicenters of the 3 largest earthquakes in 2011, M=4 to 4.7, ‘1’: 11/02/2011 event, ‘2’: 07/08/2011 event, ‘3’: 10/11/2011 event. Epicenters from the February swarm are plotted as yellow circles. Epicenters for Julian day 339 (05/12) are represented by the purple circles (cf part. 4.3.). The scale (in kilometer) is represented on left top.
3. Description and qualification of the Monasteraki borehole strainmeter

3.1. The 3-component strainmeter of Monasteraki and complementary measurements

The Monasteraki Sacks-Everston 3-component tensor strainmeter (Sacks et al. 1971) has been installed at the end of 2006 (38.403° N, 21.925° E), 2 km west to the village of Monasteraki, on the northern coast, and about 350 m north to the shoreline (MOK site). The instrument, with sensing unit of 3 m long, has been designed and constructed by the Carnegie Institution of Washington. It is cemented at 150 meters depth in a borehole within the Triassic, fracturated sandstone of the Phyllade nappe. Coring and logging of the borehole has provided evidence, within the rock mass, of thin beds of sand, as well as of sand-filled fractures, which are expected to produce an heterogeneity which might a priori alter the sensitivity of the instruments (through anelastic relaxation of strain) and/or distort it (through an anisotropic elastic response to applied stress). The strain gauges consist of six steel elliptical tubes, 3 m long, each having large ellipticity, thus compliant to traction parallel to its smallest axis, and are filled with degassed incompressible silicon oil. Diametrically opposite tubes are connected hydraulically to form a 3 component strainmeter. Volumetric strain of each gauge is transmitted through the oil to an LVDT (Linear Variable Differential Transformer) sensor, whose output is proportional to the traction. The cylinders, with vertical axis, are horizontally rotated by 120° with respect to each other, so that horizontal traction is measured in 3 directions separated by 120° (Fig. 2). This allows to resolve the horizontal stress tensor variation (amplitude and orientation of principal stress). The data logger (SOC BOX) is continuously recording at 50 Hz the three gauge signals (s1a, s1b, and s1c), with a resolution around $10^{-11}$ at short period, as well as air temperature and barometric pressure at 1 Hz.

We estimated the absolute orientation of the sensors by comparing the earth tide records to a theoretical model. The deformation of the rock mass surrounding $\epsilon_{xxx}$ in a direction $\theta$ (counted from East) normal to the gauge, due to the earth tide, can be linked to Earth tidal strain by using the relation (1) (Jaeger and Cook 1976):

$$
\epsilon_{xx} = \epsilon_{vv} \cos^2 \theta + 2\epsilon_{uv} \sin \theta \cos \theta + \epsilon_{uu} \sin^2 \theta
$$

(1)

where $\epsilon_{vv}$, $\epsilon_{uv}$ and $\epsilon_{uu}$ are respectively the east principal component, the north principal component and the nord-east shear component of solid tide.

The theoretical components of earth tide tensor in equation (1) are obtained by ETERNA 3.3 (Wenzel 1995). The orientation of each strain component could then be estimated by the angle $\theta$ giving the best waveform fit between observed traction and theoretical solid strain combination. In the theoretical tidal strain, we neglected the elastic deformation due to the Mediterranean sea tidal loading (OTL), as it was shown to be about 80 to 100 times smaller than the elastic effect of the gulf tidal loading (Canitano 2011). Further, we restricted the data analysis to a few one week long periods, where the M2 signal was much less pronounced, in order to avoid the loading influence of the nearby gulf at this tidal period. We found optimal values of $0^\circ$, $120^\circ$, and $240^\circ$ from North, for the s1a, s1b and s1c components, respectively, with an estimated uncertainty of $10^\circ$.

In complement to this recording system, down hole pore pressure in the borehole (cased down to 120 m) is monitored by a total pressure transducer (1 pt/10 min) which provides water level after barometric correction. The gulf sea level is provided by the tide-gauge installed in the marina of the Trizonia island, located about 15 km east to Monasteraki.

3.2. Instrumental noise and drift on strain records

The recorded signal on each strain gauge often exhibits clear step-like signal, lasting less than a second, as shown by the first signal step on Fig. 3 (a). These signals are due to the valve opening when the oil pressure becomes too high. The result is an extension on each gauge with amplitude around $3 \times 10^{-5}$ DU for each component. Other kind of signals are strain signatures of small fracturing process close or in the immediate vicinity of one gauge causing a mechanical stressing of the gauges by the rock. These signals could have a duration in the order of second to hour with strong amplitude, the strain shape on each component are linked to the dimensions of the fracture and its position relative to each gauge (see Fig. 3 (b)). This effect can affect one or more component.

A second characteristics of the strain signal appears when we look at longer data times series (figure 4). Each strain component exhibits a clear near linear trend. The analysis of the monthly drift indicates a fast increase of the trend value since August 2007 (first month where the instrument correctly operates), reaching a maximum one year later. Then the trend is decreasing roughly linearly.
on each component (with equivalent values for s1a and s1c) during the last three years, and end of 2011 the values are divided by three on each component (Fig. 5 (a)). This drift seems to be related to a true mechanical effect of the surrounding rocks, and not to a sensor drift (LVDT), as the latter would not produce the reported inversion of drift rate nor the correlated drift between the three sensors. A possible explanation could be the progressive tightening of the borehole forced by the lithostatic pressure, which could be due to an initially weak coupling of sensor caused by improper cementation (as suggested by water flow in the borehole revealed by temperature logging). If we considerer the 150 meters of rocks above the instrument, the lithostatic pressure results in an elastic areal strain around $10^{-4}$ by assuming standard rock properties (Poisson’s ratio $\nu \sim 0.25$, shear modulus $G \sim 30$ GPa, density $\rho_{\text{sandstone}} \sim 2.5$). This value is the same order of magnitude as the cumulative strain presently stored by each component (Fig. 5 (b)). The small difference of cumulative strain between the components could be related to a small heterogeneity close to one gauge which could perturb the signal by shear or coupling variations due to uneven cementation.

![Figure 3](https://example.com/figure3)

**Figure 3.** (a) 20 seconds of MOK strain records (50 Hz sampling). The compression is positive and the amplitude is giving in count (D.U). The steps related to valves opening are extensive with an amplitude around $3.10^5$ D.U for each component. (b) 3 hours of MOK strain records exhibiting slow strain variations related to fracturing process close to the sensor. The amplitude is giving in nanostrain (see Tab. 2 for calibration coefficients).

![Figure 4](https://example.com/figure4)

**Figure 4.** One month of MOK strain records on gauges s1a, s1b and s1c. The linear trend of each component is clearly visible, of the order of one tidal amplitude every day.

![Figure 5](https://example.com/figure5)

**Figure 5.** Strain rate (a) and cumulative strain (b) at MOK from 2007 to 2011. Components are identified by their color. The cumulative strain on each component can be related to the effect of lithostatic pressure due to the surrounding rocks.
3.3. Mechanical perturbations from external and internal sources: prediction and correction

Barometric pressure and gulf water level fluctuations are the two major strain perturbation sources in Monasteraki. At longer periods, days to months, the barometric pressure fluctuation acts in combination with the mean gulf water level and both effects are correlated. Due to the absence of rain gauge, we could not take into account the compressive signal caused by surface or infiltrated rain on land, which may produce pressure at the time scale of hours to days of the order of few millibars. We do not consider either the elastic perturbation due to the OTL from the Mediterranean Sea located 80 km west from Monasteraki, as already justified above. As the instrument is not far from the gulf (350 meters), we tested for possible sensitivity of the gauges to pore pressure diffusion from the gulf due to sea level changes. This effect is indeed very strong on the Trizonia dilatometer, located 30 m away from the gulf (Canitano 2011). For Monasteraki, a frequency band correlation between the strain signal and the water level signal does not show a significant phase shift in the case of s1a and s1c. The s1b component shows a clear phase shift, which will be discussed later, but it remains relatively small. Thus, for simplicity of a first order correction, we consider that the strainmeter is not sensitive to pore pressure diffusion effects from the sea, so that all the external forcing (sea level and atmospheric pressure) are acting instantaneously on the instrument. The only internal perturbation is the Earth tide strain, whose elastic deformation could be estimated by the relation (1) according to the gauge orientation. The overall perturbation effects are considered to be elastic, thus they combine linearly. Therefore, the estimate of the induced strain related to each forcing is done through a direct correlation between each strain component and the three forcing signals: solid tide, sea loading, atmospheric pressure. A spectral analysis of each strain component reveals the semi-diurnal effect dominated by oceanic tides and the diurnal perturbation where solid tide dominates (Fig. 6 (a)).

In addition to the perturbations above, the spectral densities exhibit clear signatures related to long period sea level perturbations. These are the mechanical effects of the free oscillations of the gulf (seiches) (Bernard et al. 2006). These oscillations with amplitude of a few centimeters are mostly triggered by wind. The s1c component seems to be less perturbed by this water mass loading, which can be explained by its orientation parallel to the coast, thus less sensitive to the expected dominantly NS traction related to local water load. The reference tide-gauge signal is recorded in Trizonia, thus 15 km from the site, so that at the period of the sea tide, the sea level at Trizonia and near Monasteraki can be considered to be in phase (Boudin 2004), and this still remains a good approximation for the 50-55 minutes seiches. At higher frequencies, the 6-8 minutes seiche period recorded in Trizonia does not appear on the Monasteraki spectral records: therefore, a low-pass filter at 40mn period has been applied to the Trizonia tide gauge signal before correlation with Monasteraki (Fig. 6 (b)).

Figure 6. Power spectral densities of the strain components at MOK (color coded). (a) Zoom at low frequency on the diurnal and semi-diurnal frequency band. The related peaks of solid and oceanic tides dominate the spectra. The s1b and s1c components are 2 times more sensitive to 24-hours period tides than s1a whereas s1c seems to be a little less sensitive (1.5 times) to 12-hours period tides than the other components. (b) Zoom at periods smaller than 1 hour. Clear peaks appear at 50-55 min and at 38-40 min period, which are interpreted as the mechanical effects of the free oscillations of the gulf (seiches).

The strain signals corrected from the linear trend and from the theoretical solid tide strain are presented in Fig. 7, and the external forcing (sea level and atmospheric pressure) are presented in Fig. 8, for June 2010. The correlation coefficients resulting from a linear combination between the three forcing signals and the strain related to each gauge are given in Tab. 1. The coefficients resulting from the analysis of a second period, 20 days long in December 2010, have been added in order to check the stability of the correlation, especially for the external forcing coupling which could be affected by a seasonal variability. The response coefficient to solid tide is stable for each gauge for the two periods, which is not surprising because the solid tide is the dominant effect in the strain signal especially at 24-hour period. The coefficient relative to the elastic effect of the sea level fluctuation also seems to be quite steady. The stability of the barometric pressure coefficient is good in the case of s1a and s1b components, but seems to fluctuate for s1c. This could be related to the stronger perturbations observed on s1c component, which perturbs the long period signal and hence the correlations with the non periodic influence of the
Table 1. Gauge response coefficients. \((C_{MS})\): solid tide \((V/nstr)\), 
\((C_{MA})\): gulf water pressure \((V/mbar)\), \((C_{PA})\): barometric pressure \((V/mbar)\).

<table>
<thead>
<tr>
<th>Gauge</th>
<th>Period</th>
<th>C_{MS}</th>
<th>C_{MA}</th>
<th>C_{PA}</th>
</tr>
</thead>
<tbody>
<tr>
<td>s1a</td>
<td>June 2010</td>
<td>49</td>
<td>81</td>
<td>366</td>
</tr>
<tr>
<td></td>
<td>Dec 2010</td>
<td>43</td>
<td>83</td>
<td>390</td>
</tr>
<tr>
<td>s1b</td>
<td>June 2010</td>
<td>52</td>
<td>50</td>
<td>353</td>
</tr>
<tr>
<td></td>
<td>Dec 2010</td>
<td>57</td>
<td>57</td>
<td>348</td>
</tr>
<tr>
<td>s1c</td>
<td>June 2010</td>
<td>57</td>
<td>56</td>
<td>559</td>
</tr>
<tr>
<td></td>
<td>Dec 2010</td>
<td>60</td>
<td>47</td>
<td>385</td>
</tr>
</tbody>
</table>

Further analysis of the barometric pressure effect thus seems to be necessary especially in the case of gauge s1c. The stability over time of the solid tide response coefficient allows us to give an accurate value for the solid calibration of each component. The calibration coefficient is simply the inverse of the tide response factor (Tab. 2). This coefficient combined with the response coefficient to each external forcing provides the induced strain due to this perturbations on each component (Tab. 3). The induced strain caused by gulf water pressure is twice smaller on s1c than in the other components, as a result of a less favourable orientation, as illustrated by the spectral densities (Fig. 6). Those also seem to argue in favour of a stronger effect on s1b than on s1c component. The s1a gauge also exhibits the stronger sensitivity to barometric pressure which may be explain by a horizontal anisotropy in response to surface load related to local elastic heterogeneity in the vicinity of the sensor.

Finally, the induced strain related to each forcing allows us to have a good prediction of the total induced strain, in agreement with the gauges orientation (Fig. 9). The predicted strain signals for June 2010 are presented in Fig. 9. The difference between the observed...
strain signal and the predicted signal gives the residual strain signal for each gauge (Fig. 10). The residual signals related to s1a and s1c gauges are relatively free from external and internal perturbations. This is not the case for s1b component, for which a 12h-period signature persists, likely associated to improper correction of the oceanic tides. This could be due to the combination of a pore pressure effect diffusion from the sea (thus out of phase with the sea level), with a partial fluid coupling of the s1b gauge due to a defect in the cementation. Thus, to improve the overall reduction of external effects, in particular at 12 hours period, a frequency-dependence of the correlations should be conducted, especially for s1b. Such a correction was made for the Trizonia dilatometer (Canitano 2011), but for simplicity will not be considered in the present paper, as for Monasteraki it is a second order correction.

3.4. Quantification of noise in residual strain records

In order to quantify the correction above, we focussed on some period range with high energy: seiches at 50-55 min, tides in the 10-14h and 22-26h period range. In the domain of the long period seiches (50-55 min), the strain residual signals are at the level of the instrumental noise (∼10^{-9}), for each component (Fig. 11). Around 12h-period (Fig. 12 (a)) the correction applied on s1a and s1c gauge allow us to eliminate respectively nearly 95% and 85% of the 12h-period perturbations, which is very satisfactory. The correction performed on the s1b gauge is less efficient, as only 65-70% of the perturbations could be removed. Around 24h-period (Fig. 12 (b)), the correction results are similar to those obtained around 12h-period, except for s1c for which the correction is slightly better here (95%). As this diurnal period is dominated by solid tide, the quality of the correction at s1a and s1c seems to argue in favour of a good estimation of the solid tide. However, the stability of the phase delay between the residuals and the observations for each frequency range, seems to indicate that the correction could be improved by taking into account a transfer function related to a simple pore pressure diffusion process from the sea.

A comparison between the strain signal related to each gauge and the water pressure fluctuation measured on the top of the bore-
The induced strain on a gauge caused by gulf water level fluctuation $G_{ij}$ (nstr/mbar) and by barometric pressure fluctuation $B_{ij}$ (nstr/mbar) retained in the present study.

<table>
<thead>
<tr>
<th></th>
<th>$G_{ij}$</th>
<th>$B_{ij}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>s1a</td>
<td>1.78</td>
<td>8.2</td>
</tr>
<tr>
<td>s1b</td>
<td>0.98</td>
<td>6.4</td>
</tr>
<tr>
<td>s1c</td>
<td>0.86</td>
<td>6.6</td>
</tr>
</tbody>
</table>

The variation of the mean noise level (in nstrain) for s1a, s1b and s1c with dominant period of observation (high-passed signal in frequency).

<table>
<thead>
<tr>
<th>Frequency</th>
<th>s1a</th>
<th>s1b</th>
<th>s1c</th>
</tr>
</thead>
<tbody>
<tr>
<td>1h</td>
<td>3</td>
<td>3.5</td>
<td>3</td>
</tr>
<tr>
<td>5h</td>
<td>6</td>
<td>13</td>
<td>7</td>
</tr>
<tr>
<td>10h</td>
<td>4</td>
<td>15</td>
<td>8</td>
</tr>
<tr>
<td>1j</td>
<td>8</td>
<td>22</td>
<td>10</td>
</tr>
<tr>
<td>2j</td>
<td>15</td>
<td>25</td>
<td>15</td>
</tr>
<tr>
<td>5j</td>
<td>20</td>
<td>30</td>
<td>25</td>
</tr>
</tbody>
</table>

...mental noise level increases with period, as suggested by Crescentini (Crescentini et al. 1997). For qualifying the detection ability for slow earthquakes observation, we assume a simple homogeneous half-space and a point source dislocation with a typical normal fault dipping 40°, and at 8 to 10 km in depth. The elastic strain is calculated at the depth of the instrument by using COULOMB 3.3, which is a revisited version of COULOMB 3.1 (Toda et al. 2005), producing, as examples, the following results. If we consider the strain threshold to observe a signal as twice the noise level, an equivalent event of magnitude 5 at 5-days period should be detected on the three components up to a maximal distance of about 15 km from the station. The two most sensitive components (s1a and s1c) should be able to detect a slow event of equivalent magnitude of 4 at 5-days period if it is located at a maximum of 3-4 km from the station. A slow event with same amplitude occurring at 8-10 km should be detected by s1a and s1c up to 1-day period.

The analysis of the residual signals during the last years does not reveal any large transient event excepted a strong signal during December 2011, lasting 5 days with an amplitude around 60-70 nstrain, most probably related to a heavy rain episode. However, during the numerous seismic crisis occurring in the rift, the strain-meter sometimes exhibits small step signals, which are presented and discussed on the next part.

4. The 2011, April 7 February seismic swarm

The western rift of Corinth has been quite active seismically during year 2011. In addition to three earthquakes of magnitude greater than 4 in the northern part of the rift, ten kilometers west from Monasteraki, some important seismic swarms have occurred...
Figure 13. Residual noise level (high-pass filtered) on two data sets (black (June 2010) and red (December 2010)). (a) 10 hours of signal at 1-hour period. (b) 4 days of signal at 10-hours period.

Figure 14. 5 days of residual noise level (high-pass filtered) at 2-days period on two data sets (black (June 2010) and red (December 2010)).

4.1. Step-like step responses

The beginning of February 2011 exhibits a high seismicity rate as indicated by the hourly seismicity (Fig. 15 (b)). The related seismic swarm is located about 5 km ENE to Monasteraki, and 5 km NW from Trizonia (Fig. 1). The activity is mostly concentrated on day 05/02 (day 36), at 5 to 8 km in depth, with epicenters clustered within 2 km. It occurred one or two kilometers east to the limit of the 22 January, 2010 Pyrgos earthquake rupture, and to its easternmost aftershocks. The strongest event had a magnitude 3.4 (05/02 at 02:52:38, NOA).

During the three days of strong seismic activity, residual strain signals in Monasteraki are exhibiting numerous small step-like signals, lasting less than a few seconds, with an amplitude of a few nanostrain (Fig. 16 (a)). Almost all of this fast variations are related to seismic events associated to the swarm, as shown by the comparison between areal strain and vertical velocity component at Pyrgos short period seismic station, located just above the hypocenter nest (Fig. 16 (b)). The largest strain variations seem to be associated to the strongest events, but some large steps are related to very small earthquakes (Fig. 16 (b), red line). The areal strain (2/3 of the sum of the three components) shows a non-random time distribution of sign, implying sequences of compressional steps, lasting several hours, alternating with sequences of dilatation. Furthermore, when these steps are recorded on 2 or 3 components, they have the same sign (95% of the cases).

We tested the hypothesis that these strain steps could be due to moment release of the coinciding seismic events, by assuming a mechanism similar to those of the nearby 2010 Pyrgos earthquakes (Fig. 1). We calculated the volumetric and areal strain for a source located at the center of the swarm, 6 km in depth, with an EW striking normal fault mechanism (strike=280°, dip=67°, rake=-90°), using COULOMB 3.3 (Toda et al. 2005). With a seismic moment equivalent moment magnitude around 4, we get an areal extension of 1 to 3 nstrain in Monasteraki. Thus, in order to associate the reported strain steps to a source at the focus of the earthquakes, the latter should have unusually large seismic moments with respect to their magnitude (between 1.5 and 3.5) deduced from their high frequency body wave radiation; alternatively, secondary sources of strain could be located much closer to the instrument, being triggered at the passage of P and/or S waves. Unfortunately, no data could be used from the Trizonia dilatometer, which was stopped in the period mid-December 2010 to mid-February 2011. Thus, in order to test these alternatives, we further analyzed the details of the coseismic strain, as follows.

4.2. High-frequency coseismic strain signals

We analyzed the 50 Hz records of the strain signals for more than 80 events (4 to 7 February), and found three main type of waveforms (Fig. 17). The first and most frequent type for all components (60-70% of the events) is a step coinciding with the arrival of the first P wave strain oscillation; the step is completed in less
than 0.5 s, so well before the S arrival about 1.5 s later. The second type (around 30% of the events) is a step coinciding with the first S wave arrival. The last type, much less frequent, shows a ramp-like signal between the P and the S, stopping at the arrival of the S waves. In each case, the related strain values are in the range 1 to 3 nstrain, with the same sign on each component for more than 95% of the cases. We also observed that only one strain event showed some step at the S arrival when it is observed at the P arrival (on s1c component), and none showed any ramp-like strain following the S wave.

We investigated the degree of correlation between the peaks of the P and S, high frequency, oscillatory strain signals, with the type and amplitude of the low frequency strain step signal. We found that when the high frequency P peak strain value is the lowest (around 0.6-0.9 nstrain for each component), the step coincides with the S wave. For larger P-wave strain, the step is usually triggered by the P waves. But the step amplitudes do not appear to be simply correlated to the high frequency wave amplitudes: P-wave induced strain of $5 \times 10^{-10}$ could sometimes trigger a step with the same amplitude than a P vibration 10 times larger. For the strain gradual increase between P and S, they appear for small values of P-wave induced strain, close to the P-step threshold value.

Figure 15. (a) Hourly seismicity during the year 2011 from the CRL-NET catalog. The red lines are indicating the 3 events with magnitude greater than 4 close to MOK. (b) Hourly seismicity during three days on February 2011 (4-7, days 35-38). The seismicity rate is dominated by a seismic swarm located about 8 km East of MOK (figure 1).

Figure 16. (a) 10 hours of residual strain signal at MOK during day 36, 2011 (5 February). The strain gauges are exhibiting some extensional or compressional steps with amplitude around 1-3 nstrain. (b) 2.5 hours of comparison between the areal strain and the vertical velocity component at Pyrgos (PYR) during day 36 (black square, figure (a)). The strain steps seem to be associated to the strongest seismic events during the crisis (figure 1). One of them (around 2325 mn, red line) is unrelated to any relevant seismic wave.

Figure 17. Example of strain variation at MOK on s1b component during seismic episodes of day 36, 2011 (5 February): (a) step after P-wave arrivals, (b) step after S-wave arrivals, (c) gradual deformation between P and S waves. In each case, the related amplitudes are around 1-3 nstrain.
We analyzed the coseismic strain signature for several events (about 15 events) from another swarm, in May-July 2009, located further away from Monasteraki, and found a similar typology, with similar amplitudes, and same sign 95% of the cases, suggesting the effect of secondary sources close to MOK.

A first interpretation would be the activation of a small fault or fracture near the instrument, triggered by the P or S vibration above some strain threshold. When the P wave does not reach this threshold, the S wave would trigger the slip. Once triggered by P or S, the slip amplitude does not depend on the amplitude of the causative waves, but on the pre-stress and friction properties of the fault, which would explain the rather stable strain step amplitude. The non-random succession of compressional or dilatational steps, as well as the stability of the strain steps, suggests the existence of either a very limited number of such secondary faults, with a persistent mechanism, or several sources sharing a similar location and mechanisms.

Under the hypothesis above, the reported absence of successive P and S steps in the same event shows that the available secondary sources are very few otherwise, for the same event, one source could provide a step on P, another one a step on S, and it would be statistically very unlikely that all would give the same response, at P or S arrival times. Then, if only a few secondary sources are active, these would have ruptured many times during the swarm sequence. Thus, as their faulting surface cannot be reloaded between successive slip events, they must have a long-term, stored shear strain that they progressively release at each slip episode. But this seems in contradiction with the absence of S step following a P step: the slip triggered by the P wave is expected to release a small fraction of the large available shear stress, thus one would expect that the following S wave could also trigger a second slip event. This last contradiction could be turned away by assuming an increase of the threshold level just after the slip due to the P waves, but no simple mechanism could support this ad hoc explanation.

Another intriguing observation is the equal sign (compression or dilatation) of the strain effect for all components in more than 90% of the cases, which strongly suggests a pore pressure effect close to the instrument, rather than slip on fractures induced by shear. Interestingly, the component with the smallest mean step amplitude, s1a, is also the less perturbed by the pore pressure diffusion from the sea, as seen previously: this suggests that a dominant cause for these steps would be pore pressure instabilities in loose cement or in sand pockets in contact with or near each gauge.

Furthermore, this model would imply that when a slip event is related to the P wave, it should produce a strain step on the three components of the instrument and similarly for the S steps. But this is not systematically observed. Furthermore, some P or S steps are mainly detected on one component: a null traction on two directions separated by 120° should occur very rarely, and requires local heterogeneity of the stress field.

Thus, from all the arguments above, a model with unstable faults near the instrument, which would slip at the arrival of P or S waves, is not likely to be the dominant cause of the reported steps. We prefer a very local effect due to some non-elastic response of the immediate vicinity of each gauge, possibly due to slip on small fractures or cracks in the sandstone surrounding the gauges, or compaction/dilatation within sand pockets or the cement itself, which could affect each gauge independently if they are at less than decimetric scale. The small size of these sources of secondary strain would explain their limited impact of the signal. However, the way that such sources could show stable sign remains to be explained.

To sum up this discussion for defining the sources of these steps, we thus reject the dominance of effects from both the seismic fault at the hypocentral location, and secondary faults near the instruments: this is not systematically observed. Furthermore, some P or S steps are mainly detected on one component: a null traction on two directions separated by 120° should occur very rarely, and requires local heterogeneity of the stress field.

As such, we thus reject the dominance of effects from both the seismic fault at the hypocentral location, and secondary faults near the instruments: this is not systematically observed. Furthermore, some P or S steps are mainly detected on one component: a null traction on two directions separated by 120° should occur very rarely, and requires local heterogeneity of the stress field.

5. Discussion and conclusion

We showed above that the Monasteraki strainmeter, in the western rift of Corinth, can provide useful, high resolution strain data for tracking slow strain transients, provided that care is taken to correct for several perturbations which will be discussed below. The long term drift, presently showing a slowly decaying rate, can be removed from the records. Its amplitude is however still too large (10^{-5} /yr) for the local tectonic strain rate to be detected, as the latter is about 10 times smaller - if this tectonic elastic strain is not relaxed by aseismic strain and creeping faults in the shallow crust. We also note that the tectonics would promote interseismic extension at MOK, thus in opposite sign with the present compressive trend.
The sensitivity to atmospheric pressure has been evaluated at the dominant period of this signal, i.e. around a few days assuming coefficient independent of frequency, which would be the case for a purely elastic shallow crust. If the latter has a non elastic behaviour a long period through some relaxation process, then there should be a frequency dependence of the sensitivity. However the noise in the data does not allow to constrain this coefficient at periods longer than a few weeks. Therefore, with the present data, we cannot exclude a long-term relaxation which could mask the tectonic loading effect, and/or slowly relax coseismic strain steps.

The strain steps related to local instabilities on fractures limit our ability to accurately evaluate the coupling coefficients between the instrument and the atmospheric and sea level loading, in the energetic period range of weeks to months. The uncertainty on these coefficients is estimated to be 10-20%. Thus, this limits the accuracy of the prediction of - and the correction for - the sea level effect on the signal at the tidal diurnal and semi-diurnal periods. The progressive reduction in the yearly count of such steps, due to the stabilization of the borehole, may allow us to improve these calibrations and corrections in the coming years. With the present correction, at periods of days to several weeks or months, between the spurious steps, the residual of the correction from these influences has an rms less than $3 \times 10^{-9}$.

The evidence for a pore pressure effect on gauge s1b, deduced in particular from a partial correction of the 12 hour sea tidal signal, is interpreted as resulting from a partially decoupled contact of the gauge with the solid rock, due to a local cementation defect. This shows that there is a significant pore pressure diffusion process between the sea and the instrument, which should be taken into account for an improved calibration procedure, considering frequency dependent, complex correlation coefficients. This is expected to further reduce the residual signal to probably less than 5% of the input signal, for all components. This unfortunately cannot be achieved yet, due to the remaining uncertainties on the sea loading coefficient, as explained above. Despite these problems, the diurnal and semi-diurnal effect of sea can be significantly reduced, as explained above. The question of reliability of strain signals from borehole, high resolution strainmeter, and of their contribution for seismic or strain transient analysis has motivated the present study. Our conclusion is that the Monasteraki, Sacks-Evertson 3-component strainmeter in the western rift of Corinth brings a very valuable contribution to the research on crustal transients, in an extensional environment, completing the continuous GPS monitoring and the Trizonia dilatometer. In particular, it reported no significant strain transient, even during seismic swarms. The resolution of the instrument, of the order of a few $10^{-9}$ at few hours and few $10^{-8}$ at few days, should be further improved in the coming years, through refined corrections from the external and internal influences taking into account pore pressure diffusion from the sea. The apparent saturation of its weak, non-linear response to high frequency seismic waves, related to very local, dynamically triggered instabilities, should be further investigated. Finally, strain transient events, if any, will be better tracked with the combined analysis of its measurements with those of the Trizonia dilatometer and of the future, strainmeters and tiltmeters planned in the area.

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