Measurement of Seafloor Deformation in the Marine Sector of the Campi Flegrei Caldera (Italy)

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Abstract We present an assessment of vertical seafloor deformation in the shallow marine sector of the Campi Flegrei caldera (southern Italy) obtained from GPS and bottom pressure recorder (BPR) data, acquired over the period April 2016 to July 2017 in the Gulf of Pozzuoli by a new marine infrastructure, MEDUSA. This infrastructure consists of four fixed buoys with GPS receivers; each buoy is connected by cable to a seafloor multisensor module hosting a BPR. The measured maximum vertical uplift of the seafloor is about 4.2 ± 0.4 cm. The MEDUSA data were then compared to the expected vertical displacement in the marine sector according to a Mogi model point source computed using only GPS land measurements. The results show that a single point source model of deformation is able to explain both the GPS land and seafloor data. Moreover, we demonstrate that a network of permanent GPS buoys represents a powerful tool to measure the seafloor vertical deformation field in shallow water. The performance of this system is comparable to on-land high-precision GPS networks, marking a significant achievement and advance in seafloor geodesy and extending volcano monitoring capabilities to shallow offshore areas (up to 100 m depth). The GPS measurements of MEDUSA have also been used to confirm that the BPR data provide an independent measure of the seafloor vertical uplift in shallow water.

1. Introduction

The Campi Flegrei caldera (Figure 1) includes the westernmost part of the city of Naples and extends into the Gulf of Pozzuoli (southern Italy, eastern Tyrrhenian basin). It is one of the most high-risk volcanic areas in the world because of its high population density and centers of significant economic activities.

Campi Flegrei is a nested caldera, formed following two catastrophic eruptions, the Campanian Ignimbrite (39 ka) and the Neapolitan Yellow Tuff (15 ka), and two successive major collapses. About half of the caldera is submerged by the sea forming the Gulf of Pozzuoli (Figure 1). Over the last 15 ka, the caldera produced tens of eruptions from monogenic vents located on land, but there is no evidence of recent eruptive vents located under the sea. The caldera is subject to successions of ground uplift generally associated with earthquakes and swarms and of seismically quiet subsidence. Between 1970 and 1984, the Campi Flegrei area showed a cumulative measured uplift of more than 3 m in the city of Pozzuoli, which is located in the center of the caldera (Del Gaudio et al., 2010). Afterward, the area underwent an almost continuous subsidence phase (Del Gaudio et al., 2010). This phase ceased in 2005 when a new general uplift phase began, which has so far reached a cumulative vertical displacement of about 43 cm (Figure 2).

The Campi Flegrei volcanic area is monitored by multiple networks of geophysical and geochemical sensors controlled by the Osservatorio Vesuviano, Istituto Nazionale di Geofisica e Vulcanologia (INGV-OV), the INGV division devoted to Vesuvius and Campi Flegrei monitoring (Figure 1). The land-based permanent monitoring system consists of a seismic network of 18 stations, a geodetic network of 17 continuous GPS (CGPS) stations, and 10 tiltmeters.

In addition, the monitoring system includes four tide gauges, a gravimetric network, a leveling network, and several geochemical stations, which together represent a complete, high-level monitoring system (Berrino et al., 2013; Chiodini et al., 2010; D’Auria et al., 2011; Giudicepietro et al., 2010).

The combination of the data collected over time by the land-based geodetic network and InSAR images, available since 1992, has contributed to the development of an interpretable model of the deformation in the Campi Flegrei volcanic area (Chiodini et al., 2015; D’Auria et al., 2015).
Figure 1. Map of the INGV-OV geophysical networks operating in the Campi Flegrei caldera. The enlarged map shows the location of the four buoys representing the MEDUSA system with the water depth at each station.
Although observations of the deformations that have occurred on land are available since the Ancient Roman era and instrumental measurements have been performed since the beginning of the last century, no measurements were available until 2008 for the submarine part of the caldera. In 2008, a first step was taken to extend the monitoring network in the submarine sector of the caldera with the CUMAS marine platform (Cabled Underwater Multidisciplinary Acquisition System) deployed at a depth of about 100 m in the Gulf of Pozzuoli (Iannaccone et al., 2009, 2010). CUMAS is a special type of buoy, which has been termed elastic beacon buoy, connected by cable to a seafloor multisensor module, equipped with autonomous power supply and real-time wireless data transmission. In order to record seismic activity and vertical seafloor displacements, the seafloor module was designed to host a broadband seismometer, a low-frequency hydrophone, and a high-precision bottom pressure recorder (BPR). The adoption of a BPR for monitoring vertical displacement was inspired by previous results on a deep underwater volcano (Chadwick et al., 2012; Nooner & Chadwick, 2009, 2016).

In late 2011, in order to have additional independent measurements of vertical seafloor deformation, a CGPS station was integrated on top of the buoy (De Martino, Guardato, et al., 2014). Three additional marine platforms, functionally similar to CUMAS but mechanically and electronically improved, were developed and deployed in the Gulf of Pozzuoli at the beginning of 2016 to constitute the new multidisciplinary marine infrastructure of four buoys named MEDUSA (Multiparametric Elastic-Beacon-based Devices and Underwater Sensor Acquisition System; http://portale.ov.ingv.it/medusa). The deployment of the three new buoys started at the end of January 2016 and was completed in about a month. After a testing period of a month, the data acquisition for the three new buoys begun at the beginning of April 2016. The CUMAS buoy suffered data transmission failures that limited the data availability.

In this paper we present the first results of the vertical displacement field of the seafloor of the Gulf of Pozzuoli obtained by GPS and BPR data recorded by the MEDUSA system in the period 1 April 2016 to 31 July 2017. MEDUSA data are compared to the vertical displacement expected in the marine sector according to the best fit Mogi model (expanding point-source) obtained by GPS land measurements only. The MEDUSA BPR data were analyzed to obtain independent measurements of the vertical seafloor uplift, which shows satisfactory agreement between BPR and GPS seafloor data. This confirms that in shallow waters, the measurement of water column pressure at the seafloor provides reliable measurements of vertical ground displacement. We conclude that GPS buoys and BPRs should be considered as core equipment for performing seafloor geodesy measurements in coastal zones subject to vertical deformation.

2. MEDUSA Infrastructure

MEDUSA is made up of four marine monitoring stations based on the CUMAS concept (Iannaccone et al., 2009, 2010). Each marine station consists of a buoy equipped with CGPS and connected by cable to a seafloor module hosting geophysical and oceanographic sensors (Figure 3a). In contrast with standard monitoring buoys, anchored on the seabed by a cable and free to move when driven by sea level changes and water currents, the floating body of each MEDUSA station is located several meters below sea level making the buoy more stable.

The stability of MEDUSA buoys is achieved with two different mechanical solutions. Buoys A and C, deployed at a sea depth of about 40 m (Figure 1), have a long steel pole inserted in the floating body and are anchored to the concrete ballast lying on the seafloor. For buoys B and CUMAS (Figure 1), located at a sea depth of about 76 and 96 m, respectively, a single long pole is impractical, and the pole is extended with a steel cable fixed to the concrete ballast.

A MEDUSA buoy, with the floating body maintaining cable tension, behaves as a semirigid system coupled to the ground providing a suitable platform for seafloor geodetic measurements: any displacement of the seafloor propagates to the emerged part of the buoy and can be thus measured by the GPS station installed.
above it (Figure 3b). As shown in Figure 3a, the buoy is equipped with a cable providing power, data transmission, and GPS clock timestamps to the seafloor module hosting the geophysical and oceanographic sensors comprising a three-component broadband seismometer, a state-of-the-art three-axis MEMS accelerometer, a low-frequency hydrophone, and a high-resolution BPR based on quartz technology. The buoy power supply system consists of rechargeable batteries connected to solar panels. The data flow from the scientific and status sensors of the MEDUSA system through 110 data channels is transmitted in real-time and in continuous mode via a 5.0 GHz radio link to the INGV-OV Monitoring Center in Naples, where they are integrated with data from the permanent land networks.

3. Data and Analysis

This section describes the analysis of data collected by the GPS station mounted on each buoy and by the BPR sensor in the corresponding seafloor module.

3.1. GPS Data

The GPS station installed on each buoy consists of a Leica GR10 receiver and an LEIAR20 LEIM antenna. The data are acquired at 30 s sampling intervals and processed in kinematic mode using the open source software package RTKLIB ver. 2.4.2 (http://www.rtklib.com) in order to take into account the movements of the buoy. The reference station chosen to process the GPS seafloor data is the LICO station (see Figure 1), located outside the Campi Flegrei caldera, at a distance of about 10 km from the caldera center (see De Martino, Guardato, et al., 2014, for a full description of the kinematic GPS processing). The GPS stations of buoys A, B, C, and CUMAS shown in Figure 1 are hereafter indicated as CFBA, CFBB, CFBC, and CFSB, respectively. Data of the first three stations are available since 1 April 2016. CFSB station suffered failures to the power system; consequently, GPS data are available in the period 7 July 2016 to 27 April 2017. The available GPS kinematic time series are given in Figure 4.

The peak-to-peak amplitudes of the horizontal components are about an order of magnitude higher than the vertical component (Up) because of the greater influence of weather and sea conditions and marine water circulation on the horizontal components. The vertical components (red data in Figure 4) have been corrected considering the vertical variation of the GPS station (dUp) induced by horizontal movements (h). It is represented by the equation \( dU_p = L - \sqrt{L^2 - h^2} \), where \( L \) is the length of the buoy (De Martino, Guardato, et al., 2014).

Here we focus on the vertical deformation GPS data only because obtaining useful horizontal geodetic seafloor displacements would require additional corrections, which are beyond the scope of this paper.

If the deformation rates are less than a few millimeters per day, the trend of the GPS measurements can be best visualized by simply averaging the displacement time series over a sliding time window (Larson et al., 2010). A weekly average is routinely applied to data produced by the permanent geodetic land network of the Campi Flegrei caldera (De Martino, Tammaro, et al., 2014; Mattia et al., 2008), and we applied...
the same averaging procedure to the vertical displacement data from the buoys using a time window of 7 days and a sliding increment of 1 day (Figure 5). This averaging procedure greatly reduces the noise in the GPS time series, and the resulting data show trends that we interpret as real uplift of the seafloor (Figure 5).

Figure 4. GPS kinematic time series computed every 30 s of CFBA, CFBB, CFBC, and CFSB stations relative to the LICO station from 1 April 2016 to 31 July 2017. The corrected vertical (Up) time series is shown in red (see text).
3.2. Bottom Pressure Recorder Measurements

BPRs have been used to measure the vertical deformation of the seafloor in deep waters at Axial Seamount (Chadwick et al., 2012; Nooner & Chadwick, 2009, 2016). The first BPR measurements of seafloor uplift in shallow water were performed using the bottom pressure time series acquired by CUMAS in 2011 at Campi Flegrei (Chierici et al., 2016) and showed a seafloor uplift of $2.5 \pm 1.3$ cm over about 1 year.

In this work we analyze the BPR time series acquired by the seafloor modules connected to buoys A and B (hereafter referred as to BPRA and BPRB) of the MEDUSA system (Figure 1) recorded in the period 29 April 2016 to 3 November 2016. This choice is due to a chunk of data showing acceptable continuity in the acquisition. Unfortunately, data from the other two BPR sensors were not available due to some technical failures.

BPR data acquired in shallow waters (see Figure 6) have to be processed to take into account the influence of tides and other oceanographic effects as well as variations in seawater temperature and salinity. In addition, the MEDUSA BPRs are subject to instrumental drift (Polster et al., 2009), and additional treatment of the time series is necessary to assess vertical displacement of the seafloor.

We combine the BPRs data with sea level data acquired from a tide gauge located in the nearby stable region of Naples following the methods of Chierici et al. (2016).

To obtain seafloor deformation at the BPRs locations, it is necessary to clean the time series from the effects of phenomena that could affect the measurements (e.g., tide, atmospheric pressure, salinity, and temperature) and to convert them to the same physical units (vertical displacement of the sensors).

We removed from the original BPR data the pressure signal generated by tidal forcing, by the variation of the water density and by the BPR
instrumental drifts. The sea level \( L(t) \) measured by the tide gauge can be described by the following equation (Chierici et al., 2016):

\[
L(t) = L_0 + ΔL(t) + \frac{ΔP_{\text{atm}}(t)}{\rho(t, T, S)g} + h_{\text{TG}}(t)
\]

where \( L_0 \) represents the average sea level (considered constant during the time of our measurements, that is, not taking into account long-term phenomena like sea level rise due to global warming etc.). \( ΔL(t) \) includes oceans waves, tides, and oceanographic components as resonances and seiches. The term \( ΔP_{\text{atm}}(t)/\rho(t, T, S)g \) describes the effect of the variation of atmospheric pressure (Wunsch & Stammer, 1997). In this term, \( \rho \) is the seawater density depending on the temperature \( T \) and salinity \( S \), and \( g \) is the acceleration of gravity. \( h_{\text{TG}}(t) \) describes the apparent sea level change due to the vertical deformation of the area (i.e., of the vertical displacement of the sensor, which is equal to zero if the tide gauge is located in a stable area as the Naples one). Similarly to the tide gauge data, the seafloor pressure data recorded by BPRs can be described by the combination of the hydrostatic load (which is dependent on the height of the column of water) and the effect due to average density of the water column caused by variation of temperature, pressure, and salinity. The changes of pressure at the seafloor \( P_{\text{bot}}(t) \) is given by

\[
P_{\text{bot}}(t) = \rho_0 g \bar{H} + \rho_s ΔH(t)g + \rho_b h_0(t)g + g \int_0^t Δ\rho(t, T, S, P) \, dz
\]

The first three terms in the second member of equation (2) represent, respectively, the hydrostatic load \( \bar{H} \) due to the average height of the water column including the atmospheric pressure; the contribution \( ΔH(t) \), due to astronomical and oceanographical components as for instance tide, ocean waves, seiches, and the tsunami and hydroacoustic waves (Chierici et al., 2016); and \( h_0(t) \) represents the vertical displacement of the seafloor due to geodynamic deformation. For each of these terms, it is necessary to consider the correct value of the seawater density \( \rho \). In equation (2), \( \rho_0 \) represents the average density of the water column, and \( \rho_s \) and \( \rho_b \) are the surface and the bottom densities of the water in the study area, respectively. In the last term of equation (2), \( Δ\rho(t, T, S, P) \) represents the variation in time of the seawater density along the water column.

As in equation (1), \( T \) and \( S \) represents the temperature and the salinity, and \( P \) is the water column pressure. Finally, \( g \) represents the gravitational acceleration. If all the components in equation (2) are known, from the BPR data can be extracted the vertical displacement of the seafloor \( h_0(t) \).

Moreover, BPRs are affected by instrumental drift, which can vary considerably from sensor to sensor and from campaign to campaign (Polster et al., 2009). The BPR drift can be described by the following (Watts & Kontoyiannis, 1990):

\[
D_{\text{BPR}}(t) = a e^{-bt} + c t + d
\]

The four parameters \( a, b, c, \) and \( d \) are dependent on the characteristics of each sensor and deployment.

We removed from the original BPR data the pressure signal generated by tidal forcing, by the variation of the water density. From the Naples tide gauge time series, we removed the effects of waves, tides, and atmospheric pressure variation.

Then we subtract the BPRs residual signal from the reference sea level time series of Napoli tide gauge to obtain a residual signal containing the vertical deformation at the BPRs locations and the BPRs instrumental drifts.

To evaluate these two contributions, we best fit the tentative deformation of the sea bottom and then the instrumental drift. Then we use the obtained drift to estimate the true sea bottom displacement using a recursive procedure (see Chierici et al., 2016, for details).

In the case of the studied uplift episode, we have used a parabolic trend to fit the deformation.

The vertical displacement at each BPR compared to the CFBA and CFBB GPS time series (same data of Figure 5) is reported in Figure 7 showing a general good agreement.

The BPR data are affected by greater noise levels than the GPS time series, with a standard deviation of 1.9 cm for BPRA and of 1.8 cm for BPRB. However, there is a general good agreement in the two trends.
The main result of this analysis confirm that BPR are suitable devices for high-resolution vertical displacement monitoring even in shallow waters, providing that the BPR signals are properly processed (see Chierici et al., 2016).

4. Modeling

The geodetic measurements acquired by the MEDUSA system have improved the monitoring network coverage of Campi Flegrei and allows to investigate the deformation of the submarine part of the caldera. To analyze the data from the GPS located on the buoys, we initially performed a best fit of a deformation model with data from GPS located on land. Next, we examine whether the best-fit deformation model determined for the land-based GPS network alone also fits the new MEDUSA data. For simplicity, we choose the Mogi (1958) model that describes the deformation at surface produced by a pressurized point-like spherical source buried in a homogeneous elastic half-space. We assumed that, during the investigated time period, the position of the source does not vary, but the strength of the source (e.g., pressure) varies with time.

The Mogi model describes the ground deformation generated by a point-like spherical cavity of radius $a$ at pressure $P$, located at depth $d$ in a homogeneous elastic half-space with shear modulus $\mu$ and Poisson ratio $\nu$. As we want to take into account the time variations in ground deformation, we assumed that pressure $P$ may vary over time and that the system responds rapidly to these variations. Under these assumptions, the radial, $U_r$, and vertical, $U_z$, components of the ground deformation are as follows (Mogi, 1958):

$$U_r(P, r) = \frac{a^3(1 - \nu)P}{\mu} \frac{r}{(r^2 + d^2)^{3/2}}$$

$$U_z(P, r) = \frac{a^3(1 - \nu)P}{\mu} \frac{d}{(r^2 + d^2)^{3/2}}$$

where $r$ is the horizontal distance from source to the point of interest.

Using vector notation, we can rewrite

$$U(P, \vec{R}) = g(P) \cdot f(\vec{R})$$

where $U$ is the generic deformation (vertical or horizontal) at position $\vec{R}$ from the center of the source, $g(P)$ is a function of the pressure, and $f(\vec{R})$ is a function of the position. As we have assumed that the pressure varies over time, we define the strength of the source $G(t)$ as

$$G(t) = \frac{a^3(1 - \nu)}{\mu} P(t)$$

where $t$ is time (see Appendix B for more details).
In order to find the best parameters for the model, we performed a best fit with the GPS data, using both vertical and horizontal components of the deformation. We assumed that the position or the depth of the center of the source does not vary over the period of time considered, whereas the strength of the source is allowed to vary with time. Despite its simplicity, the Mogi model has been widely used to model observed deformation of active volcanoes (Dvorak & Dzurisin, 1997).

The location and the strength of the source are best-fitted by minimizing the $\chi^2$ between model and GPS data (see Appendix B for details).

The land GPS data were processed using the Bernese GPS software v.5.0 in a fully automated processing chain. See De Martino, Tammaro, et al. (2014) for a description of the onshore GPS network, data recording, and processing.

Figure 5 shows a recorded uplift episode in the marine sector in the period April 2016 to July 2017. The same uplift episode was recorded by the on-land CGPS network, and Figure 8 reports the horizontal displacements (Figure 8a) and the vertical ones (Figure 8b). The time series recorded by the land CGPS stations are also shown in Figure A1. As previously recognized, the RITE station located in the center of the caldera measured the maximum observed uplift on land, in this case 7.5 cm. As shown in Figure 8b the observed uplift decreases progressively with radial distance from the RITE station toward the edge of the caldera.

The horizontal displacement pattern shows a radial symmetry centered between the RITE and CFBA stations that is consistent during this time period. The spatial characteristics of the deformation pattern suggest a single, fixed, and radially expanding source located in the Gulf of Pozzuoli, south of RITE.

The source location inferred from the data and obtained by applying the best fitting Mogi model parameters is about 0.5 km south of RITE at a depth of 2.8 km (Lat 40°49′05″N; Lon 14°07′21″E); see Figure 8.

The degrees of fit of the GPS data compared to the model are reported in Table 1.

The low root mean square (RMS) of the residual time series reported for each GPS station in Table 1 shows that the model explains the

<table>
<thead>
<tr>
<th>GPS station</th>
<th>RMS (mm; radial)</th>
<th>RMS (mm; Up)</th>
<th>N_{dat}</th>
</tr>
</thead>
<tbody>
<tr>
<td>ACAE</td>
<td>1.2</td>
<td>1.7</td>
<td>452</td>
</tr>
<tr>
<td>ARFE</td>
<td>1.4</td>
<td>1.4</td>
<td>452</td>
</tr>
<tr>
<td>ASTR</td>
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<td>260</td>
</tr>
<tr>
<td>BAGN</td>
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<td>2.4</td>
<td>467</td>
</tr>
<tr>
<td>BAIA</td>
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<td>2.2</td>
<td>464</td>
</tr>
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<tr>
<td>VICA</td>
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<td>2.4</td>
<td>466</td>
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</tbody>
</table>

Note. Root mean square of the difference between observed minus calculated displacement for the radial and vertical components. $N_{dat}$ represents the number of daily data points available for the fit.
observed horizontal and vertical components well (see also Figure A1). Next, the source of the model above was used to estimate the expected vertical deformation at the four buoy sites. These predicted values were then compared with those measured by the GPS on the buoys (Figure 9).

The RMS of the difference between observed and calculated values at each buoy is reported in Table 2. For the new buoys CFBA, CFBB, and CFBC, the RMS ranges from 3.6 to 7.8 mm, while the CUMAS buoy shows a larger RMS value of 9.3 mm. In any case, these RMS values of the difference between observed data and calculated model values are smaller than the error bars of the averaged measurements.

This comparison shows that for the studied uplift episode, the marine geodetic data are consistent with a Mogi model derived from the land-based geodetic data alone.

### 5. Discussion and Conclusions

The GPS data recorded since 2011 by the CUMAS station have measured vertical seafloor deformation in the marine sector of the Campi Flegrei caldera. During a 2012–2013 uplift episode, the seafloor followed a similar temporal pattern as that observed on land (De Martino, Guardato, et al., 2014). In 2016, three additional buoys were deployed allowing a better assessment of the deformation in the submarine sector of the caldera. Over the period April 2016 to July 2017, we performed a detailed analysis of the vertical component of the deformation recorded by the MEDUSA buoys integrating offshore data with the land-based monitoring network.

The GPS data provided by the permanent land stations (both vertical and horizontal components) were used to obtain the best-fit model parameters according to the axisymmetric model of Mogi (1958). The predicted vertical deformation pattern of the sea bottom from the best-fit model was compared with the vertical deformation measured by the MEDUSA GPS. The results show that the seafloor followed the same vertical deformation pattern observed on land and that the deformations of the aerial and submarine parts of the Campi Flegrei caldera are consistent with a unique source model. In the studied period, the maximum vertical displacement of $4.2 \pm 0.4$ cm was measured by the GPS of buoy A, which is closest to the

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**Table 2**

<table>
<thead>
<tr>
<th>GPS station</th>
<th>RMS (mm)</th>
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<tr>
<td>CFBA</td>
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</tr>
<tr>
<td>CFBB</td>
<td>4.2</td>
</tr>
<tr>
<td>CFBC</td>
<td>3.6</td>
</tr>
<tr>
<td>CFSB</td>
<td>9.3</td>
</tr>
</tbody>
</table>

*Note.* The low RMS values reported show good agreement between measured and synthetic time series.
inferred source, and the measured uplift decreased progressively with the distance of the other buoys from the source.

With the sake of outlining the perspective of sustained measures of vertical deformation at seafloor in such an important and risky area, a short discussion of the knowledge status of the Campi Flegrei ground deformation modeling and of still debated issues is valuable.

The use of the Mogi model (point source) to describe ground deformation at Campi Flegrei is not new. The strong uplift episode that occurred between 1982 and 1984 was modeled with a Mogi source by Berrino et al. (1984). Successively, also Gottsmann et al. (2006) analyzed the deformation recorded during the unrest between 1981 and 2001 by using the Mogi model. They also compared other models such as finite sphere (McTigue, 1987), penny-shaped crack (Fialko et al., 2001), and prolate spheroid (Yang & Davis, 1988). Despite the large uplift that occurred in 1982–1984 (about 1.8 m), the Mogi model reproduced satisfactorily the deformation patterns, although the authors, finally, proposed a pressurized vertically elongated prolate spheroid. More recently, Amoruso et al. (2014) showed that the deformation pattern at Campi Flegrei observed in the period between 1980 and 2010 can be decomposed into two stationary parts: a quasihorizontal elongated crack at a depth of about 3,600 m, which satisfies the large-scale deformation, and small spheroid located at about 1,900 m depth, which describes the residual deformation confined to the area of the Solfatara fumarolic field.

In this work, the analysis of data from the MEDUSA GPS shows that the deformation pattern occurred in the marine sector of the Campi Flegrei caldera is consistent with the deformation measured on land. This conclusion cannot be proved until the installation of the GPS on the four MEDUSA buoys.

During deformation episodes reported by the literature, Campi Flegrei has shown the maximum deformation in the center of the caldera, but the eruptive vents in the last 15 ka is not located in the zone of maximum uplift (Bevilacqua et al., 2015; Giudicepietro et al., 2017). Before an eruption, the radial symmetry of the deformation pattern is broken because of the intrusion of a lateral dike. This phenomenon occurred, for instance, 2 years before the last eruption episode, when an uplift of about 20 m, with a maximum located along the coast in the western part of the caldera, preceded the A.D. 1538 Monte Nuovo eruption (Di Vito et al., 2016).

We expect that in case of a nonaxisymmetric intrusion (e.g., a dike), the deformation pattern of the marine part of the caldera could provide important data for understanding the dynamics of the volcanic unrest. Apparently, when adopting a symmetric model, the seafloor measures could not add significant information. Differently, when the phenomenon requires the adoption of nonsymmetric models, the seafloor measurements are crucial and fundamental.

From this discussion, it emerges the possibility for Campi Flegrei to undergo to a lateral magmatic intrusion producing an asymmetric deformation pattern. In this framework, the availability of seafloor geodetic data is fundamental to assess the reliability of a proposed model.

As a final conclusion, we have shown that in a shallow water environment (<100 m water depth), it is possible to measure the seafloor deformation field using the network of permanent GPS and BPR such as MEDUSA. This monitoring system represents a significant achievement and advancement in seafloor geodesy. In addition, the use and validation of BPRs installed on autonomous modules on the seafloor represent complementary instrumentation for supplementing the GPS shallow water monitoring network in order to obtain denser network of high-resolution observations to address a large number of scientific applications.

Appendix A

This appendix contains the plots of the data recorded by the permanent GPS land network of Campi Flegrei and their comparison with the best-fit Mogi model. Data are referred to the period 1 April 2016 to 31 July 2017 and are averaged using a running mean with time window of 7 days and a sliding increment of 1 day. Left column shows the vertical GPS component while the radial component is showed in the right column. Radial components are computed using the position of the source inferred from the Mogi deformation model reported in Figure 8. The vertical and radial displacement computed from the Mogi model (in red) is superposed to the GPS data.
In order to model ground deformation in the volcanic area of interest, we adopted the model developed by Mogi (1958) for a point-like spherical cavity of radius $a$ at pressure $P$, located at depth $d$ in a homogeneous elastic half-space with shear modulus $\mu$ and Poisson ratio $\nu$. Moreover, to account for the time variations in ground deformation, we assumed that pressure $P$ varies over time and that the system responds rapidly to these variations (the timescale for the elastic equilibrium is much shorter than the time scale of the pressure variations). Under these assumptions, we write the radial, $U_r$, and vertical, $U_z$, components of the ground deformation as follows (Mogi, 1958):

$$U_r(P, r) = \frac{a^3(1 - \nu)P}{\mu} \frac{r}{(r^2 + d^2)^{3/2}}$$
$$U_z(P, r) = \frac{a^3(1 - \nu)P}{\mu} \frac{d}{(r^2 + d^2)^{3/2}}$$

(B1)

where $r$ is the horizontal distance from source to the point of interest and pressure $P$ is a function of time. The two formulas can be represented by

$$U(p, \vec{R}) = g(P) \cdot f(\vec{R})$$

(B2)

where $U$ is the generic deformation (vertical or horizontal) at position $\vec{R}$ from the center of the source, $g(P)$ is a function of the pressure, and $f(\vec{R})$ is a function of the position. We define the strength of the source $G(t)$ as follows:

$$G(t) = \frac{a^3(1 - \nu)P(t)}{\mu}$$

(B3)

where $t$ is time. In order to find the best parameters for the model, we performed a best fit with the GPS data, using both vertical and horizontal components of the deformation. We assumed that the position or the

Figure A1. The GPS time series data averaged over 1 week for the land stations and comparison with the best-fit Mogi model (see Appendix B). Station name is reported in each panel (see Figure 1 for station location). Left panel shows to the vertical component, and right panel shows the radial horizontal component. The red line represents the best-fit Mogi model displacement.
depth of the center of the source does not vary over the period of time considered, whereas the strength of the source is allowed to vary with time. However, the position of the source is obtained by a best-fit procedure. We express the $\chi^2$ as follows:

$$\chi^2 = \sum_{i,k} \left( \frac{(M_{ij}^k - \hat{O}_{ij}^k)}{\hat{E}_i^k} \right)^2$$  \hspace{1cm} \text{(B4)}$$

where $M_{ij}^k$ and $\hat{O}_{ij}^k$ are, respectively, the component $k$ of the deformation (vertical or radial) predicted by the model, and measured by the GPS at time $i$ and at station $j$, and $\hat{E}_i^k$ are the errors associated with the measurement of component $k$ of the deformation, at station $j$. The indices $i$, $j$, and $k$ in equation (B4) range between 1 and, respectively, the number of sampled times (e.g., days) $N_t$ the number of GPS stations $N_s$, and the number of deformation components $N_c$. In our case we set $N_c = 2$: radial ($k = r$) and vertical ($k = v$). According to our hypothesis, expressed by equation (B1), the deformation is modeled as

$$M_{ij}^k = G_i F_j^k$$  \hspace{1cm} \text{(B5)}$$

The components $F_j^k$ depend nonlinearly on both the coordinates of the source $S_x$, $S_y$, and $S_z$ and the nominal position of the GPS stations $x_j$, $y_j$, and $z_j$. Here we define the “nominal position” of the GPS station as its position in the absence of deformation. The terms $G_i$ represent the strength of the source and include the effect of the pressure, the extension of the source, and elastic characteristics of the medium. For example, for the Mogi (1958) model, $G_i$ are expressed as follows:

$$G_i = \frac{\alpha^2(1 - \nu) \rho_i}{\mu}$$  \hspace{1cm} \text{(B6)}$$

Since, without other considerations, we are not able to distinguish the contributions of the different terms (pressure, dimensions of the source, and elastic properties of the medium), we search the best $G_i$ for each time $i$.

Moreover, we assume that the spatial contributions $F_j^k$ do not vary with time but are only functions of the position of the center of the source (to be found from a best-fit procedure) and the nominal positions of the GPS stations (considered fixed in time). Note that the small variations of the position of the GPS stations from their nominal positions due to ground deformations are neglected in the terms $F_j^k$ and are considered by the terms $\hat{O}_{ij}^k$ in equation (B4). All time series for each GPS component $k$, at each station $j$, provided by the GPS processing software, contain an arbitrary offset. For consistency, we wish to have a null source strength when all the GPS displacements are zero. The offsets could be found by a best-fit procedure; however, for simplicity, we choose the first time (e.g., day) in the time series for which all the GPS are available. Then, we add an offset to time series in such a way that the new displacements for each time series are null at that time. In other words, we perform the transformation

$$\hat{O}_{ij}^k \rightarrow \hat{O}_{ij}^k - O_{ij}^k$$

where $i_0$ is the index of the selected time. After this transformation, we have $\hat{O}_{ij}^k = 0$ at $i = i_0$. Correspondingly, the best-fit procedure will produce $G_{i_0} = 0$. Actually, since the radial direction is defined only after the position of the source is defined, we set the offsets for the vertical and horizontal ($x$, $y$, $z$) components of the GPS data; i.e., we add an offset to the GPS data so that $O_{ij}^x = O_{ij}^y = O_{ij}^z$. For the minimization of $\chi^2$, we take advantage of the linearity of the terms $M_{ij}^k$ with $G_i$, as expressed by equation (B5). The minimization of equation (B4), with the use of equation (B5), is expressed by the following condition:

$$\frac{\partial \chi^2}{\partial G_i} = 2 \sum_{i,k} \left( \frac{(G_i F_j^k - \hat{O}_{ij}^k) F_j^k}{\hat{E}_i^k} \right) = 0$$  \hspace{1cm} \text{(B7)}$$

This leads to $N_t$ equations:
The terms $r_j$ represent the horizontal distance of the GPS station $j$ from the source, that is,

$$r_j = \left[ (x_j - S_x)^2 + (y_j - S_y)^2 \right]^{1/2}$$

where $x_j$ and $y_j$ are the coordinates of the GPS station $j$. The stations are assumed to lie on the horizontal plane ($z = 0$), so that we consider $z_j = 0$ for all the stations. In equation (B9) we chose a reference frame with coordinate axes $x$ and $y$ on the horizontal plane and the $z$ axis vertical and oriented upward.

For the minimization of the $\chi^2$, we search for the position of the source in the half space, inside a 3D domain where we expect to find the source, and we choose a trial position of the source. Then we evaluate the terms $F^k_j$ from equation (B9) and the parameters $G_i$ from equation (B8). These are used in equation (B5) and substituted in equation (B4).

The resulting $\chi^2$ refers to the trial source position. Then, another source position is chosen, which will lead to another value of the $\chi^2$. The optimal source will then be associated to the minimum value of $\chi^2$. For converging to the best position of the source, we used the DIRECT algorithm proposed by Jones et al. (1993). This method samples the $\chi^2$ function on a grid whose cells are automatically subdivided, in order to converge to a minimum $\chi^2$. When a minimum size of the cells is reached (1 m in our case), the search algorithm is stopped.

In summary, the procedure for finding the best position of the source, associated with the best time series of the source strength, is the following:

1. Fix the offsets of the horizontal and vertical components of the GPS data.
2. Select a 3-D search domain for the position of the source;
3. Set the source position $(S_x, S_y, S_z)$ (start with the center of the search domain).
4. Evaluate the terms $F^k_j$ ($F^H_j$ and $F^V_j$) from equation (B9) as given by equation (B10).
5. Evaluate the best source strengths from equation (B8) (linear fit).
6. Evaluate terms $\mathcal{M}^{k}_{ij}$ from equation (B5).
7. Evaluate $\chi^2$ from equation (B4).
8. Change the source position according to a specified criterion (we choose the DIRECT algorithm).
9. Repeat points 3–8 until $\chi^2$ reaches a minimum.

With the above procedure, we find the best-fitted time-varying source strength associated to the best-fitted source position for the whole time series.

The adopted procedure naturally accounts for missing data; i.e., the time series may contain days for which a subset of GPS station (or even no stations) are available.

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