Volcanic tremors: good indicators of change in plumbing systems during volcanic eruptions

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Abstract

Geophysical and geochemical signals recorded during episodes of unrest preceding volcanic eruptions provide information on movements of magma inside the lithosphere and on how magma prepares to reach the surface. When the eruption ensues continuous volcanic monitoring can reveal the nature of changes occurring in the volcano's plumbing system, which may be correlated with changes in both eruption behaviour and products. During the 2011–2012 submarine eruption of El Hierro (Canary Islands), the seismic signal, surface deformation, a broad stain on the sea surface of the eruption site, and the occasional appearance of floating lava balloons and pyroclastic fragments were the main observable signs. A strong continuous tremor in the vent accompanied the eruption and varied significantly in amplitude, frequency and dynamical parameters. We analysed these variations and correlated them with changes in the distribution of earthquakes and in the petrology of the erupting magma. This enabled us to relate variations in tremors to changes in the (i) stress conditions of the plumbing system, (ii) dimensions of the conduit and vent, (iii) intensity of the explosive episodes, and to (iv) rheological changes in the erupting magma. The results obtained show how the tremor signal was strongly influenced by stress changes in the host rock and in the rheological variations in the erupting magma. We conclude that the tracking of real-time syn-eruptive tremor signals via the observation of variations in plumbing systems and magma physics is a potentially effective tool for interpreting eruption dynamics, and suggest that similar variations observed in pre-eruptive tremors will have a similar origin.

1. Introduction

Volcanic tremors consist of long-lasting seismic signals that are often associated with the movement of fluids through a volcanic conduit in an active volcano (Schick and Riuscetti, 1973;

McNutt, 2005; Chouet, 1996; Jellinek and Bercovici, 2011). The characteristics of these tremors vary considerably between volcanoes and eruptive phases (McNutt, 1994; Jellinek and Bercovici, 2011) and depend on source effects, conduit and vent geometries, the rheological behaviour of the magma, and pressure variations in the plumbing system caused by external (tectonic and gravitational readjustments) or internal changes (magma degassing and crystallisation) (McNutt, 2005; Jellinek and Bercovici, 2011). Tremor amplitude and frequency depend on the geometry and size of the volcanic conduit, as well as the intensity (degree of explosivity) of the eruption (Chouet, 1996; McNutt, 1994; McNutt and Nishimura, 2008). The existence of further changes can be revealed by the study of other parameters based, for example, on dynamical systems theory, as in the case of volcanoes such as Stromboli (Carniel and Di Cecca, 1999), Ambrym (Carniel et al., 2003), Villarrica (Tárraga et al., 2008) and Popocatepetl (Tárraga et al., 2012). Therefore, tremors can provide information on source variations, erupting magma and/or the conduit and vent before and during the eruption (Chouet, 1996; Goto, 1999; Denlinger and Hoblitt, 1999; McNutt, 2005; McNutt and Nishimura, 2008), all of which is crucial for understanding unrest and how changes in plumbing systems affect eruption dynamics.

The 2011–2012 El Hierro submarine eruption lasted for nearly five months, from 10 October 2011 to the end of February 2012, and was accompanied by a continuous tremor signal located at the vent (Abella et al., 2012). The eruption was characterised by significant changes in the stress conditions of the plumbing system and in the rheology of the erupting magma (Martí et al., 2013a, 2013b), which in turn were reflected in changes in the location and intensity of the seismicity and the degree of crystallinity of the magma. Martí et al. (2013a) suggest that some of these changes might also have influenced some of the alterations noted in the tremor signal, although no clear cause/effect relationship has yet been established. Here we analyse the amplitude, spectral and dynamical parameters of the whole tremor signal recorded during the eruption. Our aim was to

interpret tremor variations in terms of changes in the (i) stress conditions of the plumbing system, (ii) dimensions of the conduit and vent and (iii) intensity of the explosive episodes, and (iv) rheological changes in the erupting magma at a point in time with a clear correspondence between tremor changes and variations in seismicity and/or the petrology of the erupted products. We thus identified the main tremor changes using the same type of tremor signal analysis as applied to volcanoes such as Stromboli (Carniel and Di Cecca, 1999), Ambrym (Carniel et al., 2003), Villarrica (Tárraga et al., 2008) and Popocatepetl (Tárraga et al., 2012), and compare these changes with the variations in earthquake distribution and erupting magma petrology identified by Martí et al. (2013a, 2013b). Subsequently, we establish a cause/effect relationship between tremor changes and changes in the stability of the plumbing system and/or the rheological behaviour of the magma during the El Hierro eruption. In addition, we discuss the applicability of this type of analysis to the characterisation of eruption dynamics and its potential in the interpretation of unrest episodes when applied to pre-eruptive signals.

2. El Hierro eruption

El Hierro is the youngest of the Canary Islands and its oldest subaerial rocks are dated at 1.12 Ma (Guillou et al., 1996). The island consists of a shield structure formed by different volcanic edifices and includes three rift zones along which recent volcanism is concentrated (Fig.1). The formation of El Hierro has been affected by large sector collapses that have left scars at El Golfo, Las Playas and El Julan (Fig.1). Stroncik et al. (2009) suggest that this island's plumbing system is characterised by a multi-stage ascent of magmas with major fractionation in the uppermost mantle, where incoming magma batches mingle or mix thoroughly with other ascending batches.

An unrest episode characterised by heightened seismic activity, surface deformation and gas emissions started on 17 July 2011 (López et al., 2012). Before the start of the eruption on 10

October 2011, nearly 11 000 seismic events were recorded, with local magnitudes of up to 4.3. Most hypocentres were located around 10–15 km (Fig. 1) below El Hierro's volcanic edifice (<http://www.ign.es/ign/layoutIn/sismoFormularioCatalogo.do>). The total seismic energy released before the eruption was 8.1×10^{11} Joule and the accumulated surface horizontal deformation reached 40 mm. High rates of CO₂ flux were measured in the area on land in which the epicentres were concentrated (López et al., 2012).

The eruption of basanitic magma began on 10 October 2011 at 4:35 a.m. with the opening of a N 20°-oriented fissure on the southern flank of the island at a depth of 900 m b.s.l. Over the next three days the focus of the eruption migrated northwards to a depth of 300 m, where it established a permanent central conduit and a main vent encircled by a volcanic edifice built from the accumulation of pyroclastic material. By the end of the eruption, the edifice had reached a height of 220 m with a basal diameter of over 1 km (Martí et al., 2013a). The total amount of erupted material was of the order of 0.33 km³ (Rivera et al., 2013), with an average eruption rate of 15–20 m³/s; part of the eruption corresponds to a lava flow that was emplaced in a south-westerly direction from the base of the cone. The eruption continued in a similar fashion until late February 2012, when all evidence of magma output ceased. Although the magma composition did not vary significantly, its crystal content varied from 3% during the first episodes to 44% in later ones (Martí et al., 2013b).

By comparing petrological and geophysical data Martí et al. (2013b) were able to distinguish two main eruptive episodes – each with different phases – and to identify two interconnected magma reservoirs at a depth of 20–25 and 10–15 km. The establishment of the central conduit a few days after the onset of the eruption coincided with the appearance on the sea surface of the first low-density pyroclasts and lava fragments (15 October 2011). These had a pumice-like white core of silicic composition surrounded by a black scoriaceous basanite carapace. Sigmarsson et al. (2012) identified this white pumice as a stagnant trachytic melt present as a late differentiate in the

volcanic edifice that incorporated and dissolved 10-15% of the quartz sand – heated and remobilized by the basanitic magma – present on the sea floor below El Hierro. Later samples were basanitic with no contamination by silicic material (Martí et al., 2013b), thereby indicating that the formation of the white pumices was an anecdotic part of this eruption.

After 10 days of very weak seismicity, strong (volcano-)tectonic seismicity occurred, centred in the north of the island mostly at depths of 20–25 km but also 10 days later at 10–15 km (Martí et al., 2013a). In early November new intense seismicity was detected in the north at a depth of 20–25 km. By the end of November seismicity had decreased drastically and was mostly concentrated at a depth of 10–15 km, first in the north but later also in the south. This coincided with a marked change in the rheology of the erupted magma, which became much less crystalline. This was interpreted by Martí et al. (2013b) as the result of a new deeper magma intrusion into the shallower reservoir at the beginning of the second eruptive episode, in which hotter and more fluid basanitic magma was expelled.

In addition to the (volcano-)tectonic seismicity, the eruption was accompanied from its outset by a continuous strong tremor in the vent linked to the movement of fluids in the upper part of the volcanic conduit (Martí et al., 2013a).

3. Methodology

Albeit at different intensities, the tremor signals were recorded by all of the seismic stations of the IGN monitoring network (López et al., 2012; Fig. 1) during the eruption. For the purposes of this study we analyzed the vertical component from all of these stations. To describe the time evolution of the features of the tremors a data reduction was carried out by computing a selection of parameters that contain most of the information content hidden in the signal (see Carniel and Di Cecca, 1999). The tremor ground motion amplitude (in $\mu\text{m/s}$) was computed by averaging the

modulus of the signal. We subdivided the raw seismic data into windows of 60 s (i.e. 6,000 sample points at 100 Hz) with no overlap. In each window, we computed the amplitude as the integral of the modulus of the signal for any single component after removing the instrumental offset.

Next, we computed the time evolution of the spectral content for the same time windows using a data reduction procedure (Carniel et al., 2003), depicted in Figure 2 for three days at the beginning, three days during and three days at the end of the eruption. The part of the spectrum below 10 Hz was divided into 40 bins and the resulting values were normalized to examine the relative distribution of the energy across this frequency range but not its absolute values (the absolute variations are examined by the tremor intensity). Two scalar parameters derived from the normalized spectrogram – the average frequency and the dominant frequency – summarize the frequency content. For each time window, the average frequency is the average of the bins' central frequencies weighted by their own spectral value (Carniel et al., 2003); this basically represents the centre of mass of a given spectral distribution. The dominant frequency is the central frequency of the bin in which the maximum spectral value is reached. In order to highlight significant changes, the graphs are plotted after applying a 60-minutes moving average.

The other parameters analyzed were derived from dynamical systems theory, a methodology that has already been applied successfully to volcanoes such as Stromboli (Carniel and Di Cecca, 1999), Ambrym (Carniel et al., 2003), Villarrica (Tárraga et al., 2008) and Popocatepetl (Tárraga et al., 2012). In the case of El Hierro, we analysed the time evolution of two dynamical parameters: the first zero of the autocorrelation function (Fraser and Swinney, 1986) and the singular values produced by the Singular Value Decomposition (SVD) (Broomhead and King, 1981) (see Appendix for a detailed description). Each of these parameters provides an independent piece of information; some transitions are seen much more clearly in one or just a few of these parameters (Carniel and Tárraga, 2006; Carniel et al., 2006). Figure 3 shows the time evolution of several parameters

derived from the continuous seismic signal as recorded at the CMCL station (Fig. 1) in October. Classical parameters – the average amplitude filtered from 1 to 10 Hz (Fig. 3a) and the frequency content (Fig. 3b, c) – are shown, along with the dynamical parameters: the first singular value (SV) (Fig. 3d), the r_k ratio from the SVD (Fig. 3e), and the first zero of the autocorrelation function (Fig. 3f). This initial period of the eruption was the most significant in terms of variations in the dynamical parameters, which were to become less relevant during the rest of the eruption.

4. Time evolution of tremor features

Figure 4 shows the evolution over time from 1 October 2011 to 29 February 2012 of the whole seismicity (Fig. 4a–d), and of the average amplitude (Fig. 4e) and the average (Fig. 4f) and dominant (Fig. 4g) frequencies of the seismic signal. At the beginning of the October, the amplitude was close to 0.5 $\mu\text{m/s}$, with occasional peaks not exceeding 2 $\mu\text{m/s}$. On 10 October 2011 the onset of the tremor took the form of a sudden increase in the amplitude that reached a maximum value of 8 $\mu\text{m/s}$ on 11 October 2011, before dropping suddenly on 12 October 2011. After this decay the amplitude varied in the range 2–3 $\mu\text{m/s}$ until early November. On 3 November 2011 there was another sharp increase and the amplitude reached values close to 7 $\mu\text{m/s}$. During the subsequent 15 days the amplitude remained relatively high with values of 4 $\mu\text{m/s}$ but reached 10.3 $\mu\text{m/s}$ on 18 November 2011, the maximum amplitude of the whole eruption tremor. From 20 November 2011 onwards the tremor began to decrease gradually to the end of January, interrupted only by occasional increases (in late November, from 2 to 4 $\mu\text{m/s}$; in the middle December, from 1 to 3 $\mu\text{m/s}$; and in late January from 1 to 5 $\mu\text{m/s}$). Subsequently, the amplitude decreased progressively until the end of February, at which point it returned to its pre-eruptive values.

The dominant frequency before the onset of the tremor varied greatly, with values in the range 0.4–4.5 Hz. From the tremor onset until the end of October the dominant frequency was much more

stable, with values mostly in a low frequency range of 0.5–1.2 Hz. Between the end of October and 20 January 2012 an almost constant value of 1 Hz was observed, with only occasional peaks of 1.8–1.9 Hz. From the end of January onwards the dominant frequency varied in a range of 0.4–1 Hz, with two peaks of 3 and 4 Hz on 25 and 28 January 2012, respectively (Fig. 4g). In February 2012 the dominant frequency values oscillated between 0.4 and 0.8 Hz but by the end of the eruption the maximum values resembled the values recorded at the onset of the eruption. The average frequency showed significantly different behavior before and after the tremor onset, with values in the range 6–10 Hz before the onset and lower values of 2–5.5 Hz after the start of the tremor. By around 20 January 2012 these values oscillated in the range 3–8 Hz but by February 2012 were similar to the pre-eruptive values.

5. Discussion

The characteristics of the tremor on El Hierro suggest that it originated mainly in the uppermost part of the eruption conduit and so the variations in the continuous seismic signal should be in part due to changes in this section of the conduit or at the vent. However, Martí et al. (2013a) suggest that certain changes observed in the tremor signal could correlate to changes in deeper parts of the plumbing system and also to variations in magma rheology during the course of the eruption. The main variations in the plumbing system and in magma rheology can be associated with changes in recorded geophysical parameters (e.g. seismicity and surface deformation) and in the petrology of the erupted products, which have been documented by Martí et al. (2013a, 2013b) for the El Hierro eruption. Unfortunately, given that this eruption could not be observed directly and that samples of the erupted products were only available from a few days after the start of the eruption the record of the eruption and its different phases and pulses is incomplete. However, the available information is sufficiently robust to allow for correlation with the tremor signal, at least, that is, for those time

windows for which geophysical and petrological information exists.

At 4.35 a.m. on 10 October 2011 the continuous seismic stations underwent a sudden increase in amplitude (Figs. 3a, 4e) with the appearance of a volcanic tremor signal pattern, indicating that the eruption process was commencing. Initially, the vent consisted of a narrow eruptive fissure that in the first three days of the eruption moved nearly 5 km northwards (Martí et al., 2013a) from a depth of 900 m to 300 m. Although the tremor signal fluctuated somewhat during this period, its amplitude remained high and the tremor onset is best observed in other parameters: for example, the average frequency (Figs. 3b, 4f) decreased abruptly, while the dominant frequency (Figs. 3c, 4g), which varied greatly before the tremor onset, became much more stable, above all in the low frequency range. The first SV (Fig. 3d) varied quite erratically before the tremor onset (values in the range 4–13) but behaved much more coherently afterwards. The r_k ratio of the SVD (Fig. 3e) shows a sharp transition and suddenly increased by a factor of 5. The first zero of the autocorrelation function (Fig. 3f) shows another clear transition, dropping abruptly from a value of 33 to a value of 15 on 10 October at 4.35. Although a significant change is also seen in the time evolution of the spectral and amplitude parameters, the transition in the first zero of the autocorrelation and in the r_k parameter are much more evident and abrupt (Fig. 3).

The northwards progression of the eruptive fissure finished at the end of 12 October 2011 when it intersected a E–W-oriented normal fault, thereby creating the central conduit and vent that were maintained until the end of the eruption (Martí et al. 2013a). At this point, the amplitude of the tremor signal decreased abruptly to about 80% (Figs. 3a, 4e). We interpret this first major change in the tremor amplitude to be the result of the variation in the flow rate that occurred when the eruption changed from a fissure to a central conduit (Ferrazzini and Aki 1992; Konstantinou and Schlindwein, 2002). Nevertheless, this change in the flow rate did not affect to the same extent either the frequency content or the dynamical parameters (Fig. 3).

The situation remained more or less stable until 15 October 2011, when a new but minor increase in the tremor amplitude – probably indicating an increase in the eruption explosivity – was observed (Fig. 3a). However, the r_k ratio of the SVD showed an abrupt, tenfold increase in relation to the pre-eruptive signal (Fig. 3e). This change corresponded with the appearance on the sea surface of the first erupted products. The high vesicularity of the mingled pumice fragments that were sampled could explain the increase in the tremor amplitude as the product of an increase in the gas content of the erupting magma caused by the incorporation of trachytic material and the assimilation of silicic sediments. The tremor amplitude (Fig. 3a) – and more obviously, the dynamical parameters (Figs. 3d–f) – decreased drastically again on 17 October 2011, day on which these pumice fragments ceased to appear on the sea surface. This probably indicated the end of the magma mixing and assimilation episode and thus a decrease in the gas content in the erupting magma and its explosivity.

Strong seismicity started around 22 October 2011 in the north of the island at a depth of 20–25 km (Figs. 4a–d). The first appearance during the eruption of seismicity at deeper levels was accompanied by changes in the tremor signal. Although the amplitude and frequency did not change significantly (Figs. 3a–c), there were visible changes in the other parameters such as r_k (Fig. 3e) and in the first zero of the autocorrelation function (Fig. 3f). Martí et al. (2013a) have interpreted this deeper seismicity as a result of a tectonic and gravitational readjustment of a deeper reservoir following decompression after the first few days of the eruption. In this case, the variation observed in the tremor signal would seem to suggest that the seismic tremor located at the vent was reacting to events occurring more than 20 km away to the north.

During the period 3–7 November 2011 a new increase in the amplitude of the tremor signal was observed (Fig. 4e), coinciding with clear evidence of an increase in the eruption intensity (a giant stain and occasional giant bubbles observable on the sea surface above the eruption site).

On 16 November 2011 there was a spectacular increase in the tremor amplitude, which reached its highest values since the beginning of the eruption. However, on 18 November 2011 a sharp decrease in the amplitude was noted, followed 48 hours later by another sharp increase (Fig. 4e) that was apparently not related to any change in the distribution of the earthquakes (Figs. 4a–d). The lack of any samples of the erupted products from that day precludes establishing any correlation with possible petrological changes. However, an increase in the intensity of the eruption due to a rise in the gas content of the magma and/or to a narrowing of the conduit would seem to be the most plausible explanations, since the tremor signal followed a similar pattern to that observed at the beginning of the eruption. This episode could have been associated with the appearance of secondary vents on the volcano – detected by a bathymetric survey conducted a few days later (Rivera et al., 2013) – and with the origin of the aligned circular stains that appeared on the surface of the sea during the eruption.

Around 20 November 2011 the tremor amplitude decreased drastically and remained low until the end of the eruption. This new major change in the amplitude of the tremor was probably associated with the transition between the two eruptive episodes during which the plumbing system was recharged with less viscous, hotter and less gas-rich magma, as was determined from the samples collected a few days later (Martí et al., 2013b). Thus, we suggest that changes in the composition of the erupting magma, which could imply changes in its rheology and gas content, may also change the characteristics of the tremor signal.

Around 29 December 2011 there was a decrease in amplitude, which was not related to any compositional change in the magma but rather to a change in the depth and location of the seismic events (Fig. 4b, 4c). We observed a significant change in the dominant and average frequencies (Figs. 4g, 4f) and amplitude (Fig. 4e) of the tremor signal around 20 January 2012, which coincided with another increase in the occurrence of earthquakes and a slight increase in the crystallinity of

the samples collected a few days later. This suggests once again that variations in the viscosity of the erupting magmas may lead to observable changes in the tremor signal.

Approximately between 20 January and 14 February 2012 the tremor signal was characterized by pulsations and oscillations in the dominant frequency (Fig. 4g). These oscillations are best-seen in the spectrogram (Fig. 2b). We believe that this pulsating form of tremor corresponds to the transient dynamics of the final few days of the eruption and are probably associated with a discontinuous supply of highly degassed magma that could not sustain a continuous flow into the vent.

Finally, in the rest of the tremor signal the dominant frequency of the seismic signal decreased to the same level as before the start of the eruption (Fig. 2c, Fig. 4g), thereby suggesting that in the waxing phase of the eruption no other changes in the plumbing systems occurred.

6. Conclusions

Given that the characteristics of the tremor signal are highly influenced by stress changes in the host rock and by rheological variations in the erupting magma, our results suggest that tracking real-time tremor signals may provide an effective tool for describing eruption dynamics in terms of plumbing system behavior and magma physics. The fact that an eruptive tremor originating at the vent responds in real time to changes in the plumbing system occurring at a distance of over 20 km suggests that any pressure (stress) change affecting the magmatic system is immediately transmitted to all parts of magma, thereby demonstrating the interconnection of the whole plumbing system during the eruption and the dependency of eruption dynamics on its behaviour. In this study, we aimed to interpret changes observed in the tremor signal that are clearly correlated in time with changes in seismicity and petrology. Moreover, we were able to identify many other changes that do not obviously correspond to variations in the distribution of the seismicity. Nevertheless, they could

be related to petrological or rheological changes that were not recorded in the samples studied. For example, small variations in the degree of crystallization corresponding to changes in magma degassing could have caused some of these minor variations in the tremor signal and, albeit a purely speculative hindsight, the appearance of similar patterns does suggest similar causes. Although the inherent limitations of our observations preclude a complete interpretation of the tremor signal in terms of eruption dynamics, they do not invalidate the main conclusion of this study, namely, that the tremor signal is strongly dependent on the behavior of the plumbing system during the eruption and that the correct interpretation of the syn-eruptive tremor signals is crucial for understanding eruption dynamics.

Our results also suggest that, although classical parameters that quantify intensity variations and spectral content are able to highlight significant transitions in the eruption dynamics, the use of additional, more sophisticated parameters based on the theory of dynamical systems provides a better characterization of eruptive phases.

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Appendix

The dynamical approach to time series analysis is based on the fact that a time series can be considered as the observable part of a *dynamical system* that underlies the phenomenon under study. This idea was developed in a series of seminal papers in the 1980s (Packard et al., 1980; Takens, 1981; Eckmann and Ruelle, 1985). A volcano can thus be modeled as a continuous dynamical system, whose state is represented by a *state vector* of variables that evolve with time in a *state space* according to the set of equations that completely determines its behaviour (Carniel and Di Cecca, 1999).

The aim is to reconstruct an m-dimensional state from a mono-dimensional time series using a time delay. In our case the mono-dimensional time series is the vertical ground velocity. In order to reconstruct this multidimensional trajectory we first need to determine the delay time and the embedding dimension m.

The simplest method for choosing the delay time is to use the autocorrelation function $\rho_{XY} = \sigma_{XY} / \sigma_X \sigma_Y$, where σ_{XY} represents the covariance and σ_X and σ_Y the standard deviation of

the random variables X and Y , respectively. We can thus define $X = \{x_k\} \ k = 0, \dots, N-\tau$ (the original time series) and $Y = \{y_{k+\tau}\} \ k = 0, \dots, N-\tau$ and $\tau = 0, \dots, \tau_{max}$ (the delayed time series), where N is the length of the time series. After computing the correlation between the original time series and the delayed time series for values of τ from 0 to a maximum delay τ_{max} , the optimal delay time will be the first for which ρ_{xy} is zero, i.e. the first zero of the autocorrelation function. This choice guarantees linear independence of the coordinates. In this work, we examined the temporal evolution of the first zero of the autocorrelation. For the same time windows that were used for the spectrogram and after removing the offset of the signal (by subtracting the time average over its duration), the autocorrelation function was computed for every τ between 1 and 20, the first zero-crossing was determined and finally an hourly average was computed. This parameter is related to the spectral content of the analyzed signal. For pure sinusoidal signals, the lower the frequency, the greater the optimal delay time. As shown in some examples presented here, this dependence, nevertheless, is not quite so straightforward for experimental signals such as volcanic tremors. The stability/variation over time of this parameter – if added to a classical spectral analysis – provides us with an additional indication of the existence of different volcanic regimes and/or the recognition of a transition between them.

The embedding dimension m is the other parameter that is required for a multistate reconstruction and we determined it using the SVD method (Aldrich and Barkhuizen, 2003). The one-dimensional time series is first recast as an L -dimensional time series known as a *trajectory matrix* associated with a window length that coincides with the embedding dimension. The series is expanded, on this occasion with a unitary lag, to create a number of lagged vectors. For a time series $\mathbf{x} = \{x_1, x_2, \dots, x_n\}$ and an embedding dimension m , the resulting trajectory matrix is

$$\mathbf{X} = \begin{bmatrix} x_1 & x_2 & \cdots & x_m \\ x_2 & x_3 & \cdots & x_{m+1} \\ \vdots & \vdots & \ddots & \vdots \\ x_{n-m+1} & x_{n-m+2} & \cdots & x_n \end{bmatrix}$$

In the second step (which is a Singular Value Decomposition – SVD – proper), the trajectory matrix is decomposed as $\mathbf{X} = \mathbf{U} \mathbf{S} \mathbf{V}^T$, where $\mathbf{U}$ is a $(n-m+1) \times (n-m+1)$ orthogonal matrix, $\mathbf{V}$ is an $m \times m$ orthogonal matrix and $\mathbf{S}$ is a $(n-m+1) \times m$ diagonal matrix whose elements are the singular values of $\mathbf{X}$. This is computed using the covariance matrix $\mathbf{C} = \frac{1}{n-m} \mathbf{X}^T \mathbf{X}$ and determining a spectral decomposition of this matrix $\mathbf{C} = \mathbf{\Phi} \mathbf{S} \mathbf{\Phi}^T$, where $\mathbf{\Phi}$ is an orthonormal matrix whose columns are the eigenvectors of $\mathbf{C}$, and $\mathbf{S}$ is a diagonal matrix in which the elements σ_i^2 are the eigenvalues of $\mathbf{C}$ in decreasing order.

The singular values of $\mathbf{X}$ are given by the square root of the eigenvalues of $\mathbf{C}$. Therefore,

$$\mathbf{S} = \begin{bmatrix} \sigma_1 & 0 & \cdots & 0 \\ 0 & \sigma_2 & \cdots & 0 \\ \vdots & \vdots & \ddots & \vdots \\ 0 & 0 & \cdots & \sigma_m \\ 0 & 0 & \cdots & 0 \\ \vdots & \vdots & \ddots & \vdots \\ 0 & 0 & \cdots & 0 \end{bmatrix}, \sigma_1 \geq \sigma_2 \geq \dots \geq \sigma_m \geq 0$$

In this way we are able to determine a vector base in which the vectors are ordered in terms of their importance in the construction of the signal. This importance is given by the weight σ_i of each singular value of the matrix $\mathbf{X}$. We can also weight the importance of a given (low) number k of first

dimensions and monitor how this low-dimensional subsystem gains or loses importance with respect to the remaining $(m-k)$ dimensional space. This in turn can help highlight the fact that the system tends to become simpler. Numerically, this corresponds to the calculation of the ratio

$$r_k = \sum_{i=1}^k \sigma_i / \sum_{i=k+1}^m \sigma_i$$

for which we can then analyze the time evolution one again. Broomhead and King

(1986) suggest that the choice of the size m of the windows should be evaluated by taking into account the resulting window length $\tau_w = m\tau_s$. These authors propose a number of bounds for the choice of m , indicating that the resulting window length τ_w should verify $(2f+1)\tau_s \leq \tau_w \leq \tau^*$, where $\tau^* = 2\pi/\omega^*$ is the band-limiting frequency, assuming that the time series does not contain frequencies with significant powers greater than this cut-off frequency. However, the dimension f of the embedding manifold is unknown and for an experimental time series the sampling period τ_s is already determined; for this reason in all the method of delays it is commonest to concentrate on the choice of m [Packard et al., 1980]. Both previous examples of an analysis of an experimental time series of a volcanic tremor [Carniel and Di Cecca, 1999] and numerical experiments [Bozzo et al., 2010] suggest that $m=20$ is a reasonable choice, which is the figure we adopt in this study. Nonetheless, it is the time evolution of the singular values that yield additional information about statistically significant changes in the volcanic system under study; we are therefore interested only in *relative* variations of the SVD-related parameters and the window size is not a critical choice as long as it is reasonable.

Figure captions

Figure 1. DEM of El Hierro showing the main structural and morphological features, the IGN seismic network and the distribution of seismic events recorded from 10 October 2011 to the end of February 2012 (around 3500 events).

Figure 2. Example of waveforms and spectrograms of the seismic signal recorded at the CMCL station during (a) three days at the beginning of the eruption, (b) three days during the eruption and (c) three days towards the end of the eruption.

Figure 3. Example of the analysis of the tremor signal recorded during October 2011. Time evolution of (a) average amplitude, (b) average and (c) dominant frequencies, (d) First SVD, (e) ratio r_k and (f) delay time suggested by the first zero of the autocorrelation function of the continuous seismic signal. Vertical lines show the date of the main changes observed in the tremor signal (see text for further explanation).

Figure 4. Time evolution of the earthquakes on El Hierro with $M \geq 2$ and tremor features from 1 October 2011 to 29 February 2012: (a) Magnitude, (b) Depth, (c, d) Epicentral coordinates (latitude; longitude), (e) Seismic signal amplitude, (f) Average frequency and (g) Dominant frequency. Vertical lines show the date of the main changes observed in the tremor signal (see text for further explanation).

Figure 1

Figure 2

Figure 3

Figure 4

Highlights

- Time evolution of spectral and dynamical parameters is computed.
- Changes are associated to stress and rheology variations.
- Syn-eruptive tremor is an effective tool to interpret eruption dynamics.