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Key Points:
- Unusual signals, with monochromatic Rayleigh wave trains of long duration, are attributed to the transit of lava in a shallow opened conduit
- Under a high hydrostatic pressure of ~400 bars, the transit of lava excited the resonance of this conduit
- Such identical seismic events occurred on 27 May 27 2011 and 29 April 2013 at the submarine volcano Rocard, Society hot spot

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Unusual seismic activity in 2011 and 2013 at the submarine volcano Rocard, Society hot spot (French Polynesia)

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Abstract We analyze two seismic events that occurred on 27 May 2011 and 29 April 2013 at the Rocard submarine volcano which overlies the Society hot spot. The Polynesian Seismic Network recorded for the first time unusual associated short- and long-period signals, with perfectly monochromatic (0.0589 Hz) Rayleigh wave trains of long period and duration. None of the numerous observations of long-period (10–30 s) signals previously associated with volcanic activity in Japan, Italy, Mexico, Indonesia, Antarctica, and the Hawaiian Islands have the characteristics we observed at Rocard. We propose a tentative model for these unusual and rather enigmatic signals, in which the movement of lava excited the resonance of a shallow open conduit under a high hydrostatic pressure of ~400 bars.

1. Introduction

Volcanic eruptions are preceded and accompanied by seismic signals that sometimes allow volcanologists to forecast these eruptions and to analyze the mechanisms of magma ascent below and within volcanic edifices. These signals are especially useful for anticipating eruptions of active oceanic hot spot-related shield volcanoes such as Kilauea (Hawaii), Piton de la Fournaise (La Réunion), and several volcanoes in Iceland. The aim of this paper is to describe highly unusual seismic signals recorded in 2011 and 2013 by the Polynesian seismic array (Réseau Sismique Polynésien hereafter RSP). Their origin is traced to the active submarine volcano Rocard overlying the Society (Tahiti-Mehetia) hot spot region [*Cheminée et al.*, 1989; *Binard et al.*, 1991]. These short- and long-period signals differ considerably from those during previous seismic crises at other active submarine volcanoes in the Society Hot spot region, e.g., Mehetia in 1981 and Teahitia seamount in 1982, 1983, 1984, and 1985 [*Talandier and Okal*, 1987].

The RSP network consists of several subarrays of short-period seismometers deployed on various linear island chains in French Polynesia; the locations of the Tahiti and Mehetia stations which are the closest to the Rocard volcano are shown in Figure 1. This network is complemented by several broadband seismometers. The short-period seismic waves were assumed to be *P* waves, whereas the long-period wave train was assumed to be Rayleigh wave, according to its retrograde elliptical particle motion in the vertical plane, in the direction of propagation. For each event described here, two independent locations were determined from the data of short (Pb and Pn) and long periods (Rayleigh), and an additional location was determined by mixing the short- and long-period arrival times. The best results are obtained for a velocity of Rayleigh waves of 3.6 km/s, which is in agreement with a period of 17 s for the propagation model of the Tuamotu Archipelago [*Okal and Talandier*, 1980].

The relative large error ellipses of the six epicenter locations (Figure 1, inset) is due to (i) the poor geometry of the RSP network with respect to the epicenters, (ii) the uncertainty on the arrival times of the short-period volcanic tremor, and (iii) those of the long-period Rayleigh waves. These epicenters are clearly located on or near the submarine Rocard volcano, but the exact location of the sources with respect to the edifice cannot be determined. The locations of the sources of volcanic tremor and Rayleigh waves cannot be differentiated, and so we interpreted that the 2011 and 2013 events may be due to the same sources.

2. Characteristics of the Signals

The short-period signals consist of Pb and Pn waves of fluctuating amplitude which lasted, respectively, from 40 to 50 min on 27 May 2011 at 22:17 UT and from 30 to 40 min on 29 April 2013 at 21:06 UT. In both cases,

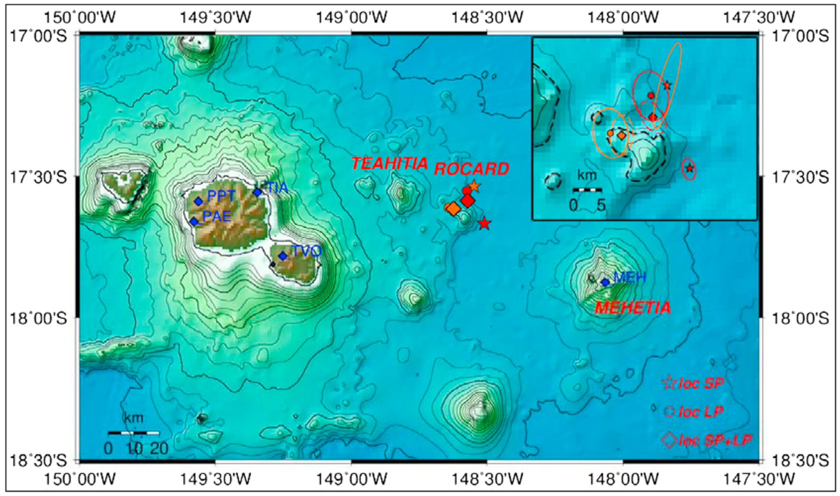

Figure 1. Map showing the locations (in blue) of the seismic stations on three different volcanic edifices of the Tahiti-Mehetia hot spot (Mehetia MEH, Taravao TVO, and Tahiti TIA, PPT, and PAE) and the epicenters deduced from the data of (i) short-period events (Pb, Pn) shown as stars, (ii) long-period events (Rayleigh) shown as circles, and (iii) associated events of both short and long periods shown as diamonds. The inset shows the confidence ellipses (red: 2013; orange: 2011) which frame the submarine volcano Rocard but do not provide a precise location of the sources with respect to the edifice. Thick lines denote the −3300 m isobath.

the corresponding spectra are rich in low frequencies with maximum energy between 0.5 and 4 Hz (Figure 2). This low-frequency volcanic tremor is associated with rare higher frequency signals from impulsive events. The long-period signals consist of perfectly monochromatic Rayleigh wave trains that lack the usual frequency dispersal characterizing Rayleigh waves of oceanic propagation (Figure 3). They were received at PPT (Society Archipelago), TBI (Austral Islands), RAR (Cook Islands), and RKT (Gambiers Islands) seismic stations. The durations of these wave trains were independent of the distance from the source, i.e., the Rocard submarine volcano, located at 100 km from the PPT station but at 650, 1250, and 1560 km from the three other stations. The frequency of the wave train was 0.0589 Hz (17.0 s) and was identical in 2011 and 2013 (Figure 3).

The main features of the long-period signals in 2011 and 2013 are shown in Figures 3 and 4. The only differences between them are (i) their respective durations, 50–60 min in 2011 versus 30–40 min in 2013, (ii) their amplitudes, approximately 20% weaker in 2013, and (iii) the occurrence of two partly overlapping sequences

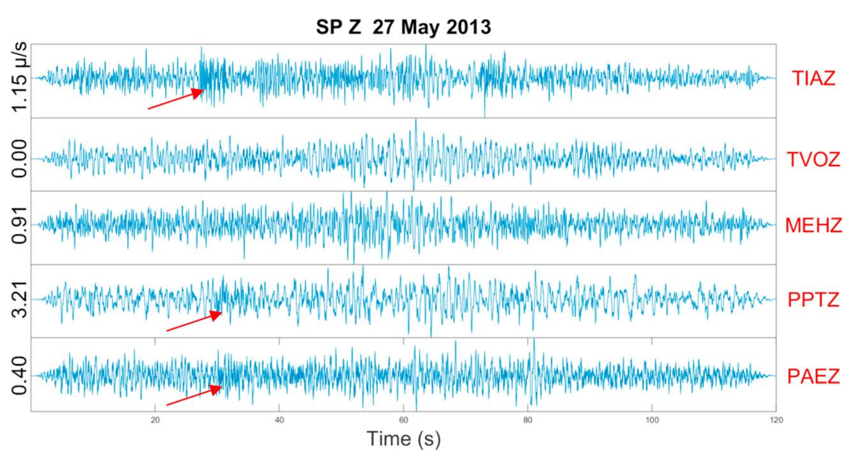

Figure 2. Sequence of short-period volcanic tremor consisting of Pb and Pn waves of fluctuating amplitude at Rocard volcano. The red arrows point to parts of the signals with high-frequency volcanic tremor that are missing from the stations of Mehetia (*MEH*) and Taravao (*TVO*). The location of the 2013 epicenters was determined by data received by the Rangiroa network at a distance of 200 km, which also recorded the Pn waves of these rare impulsive events.

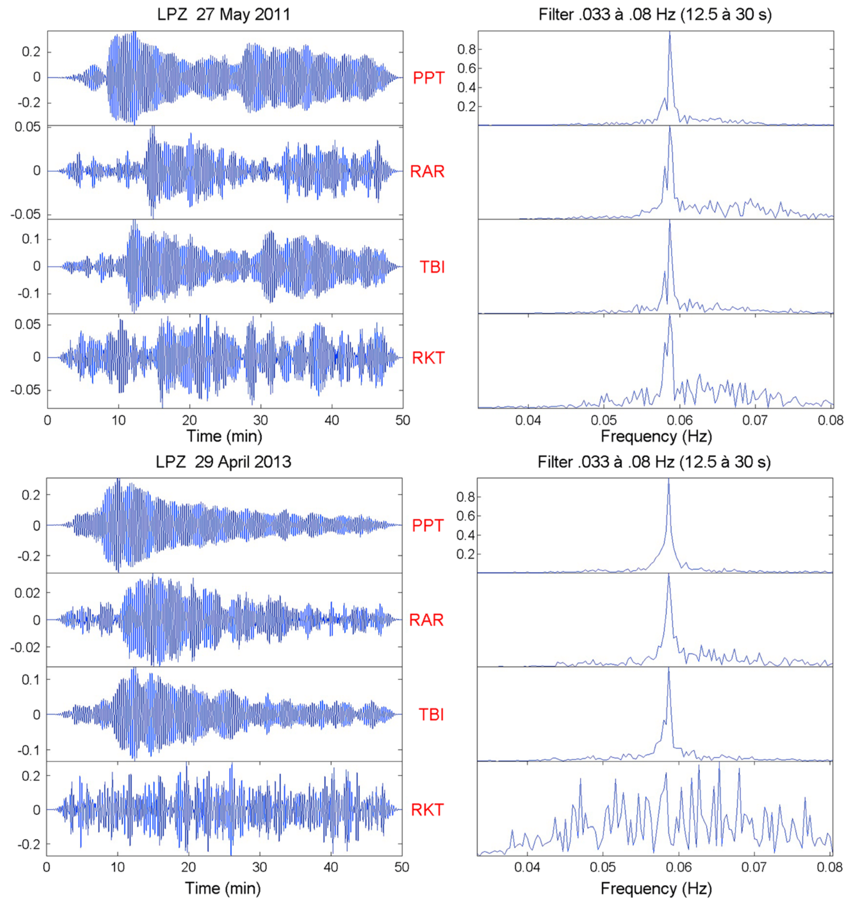

Figure 3. Rayleigh wave trains of long period visible at PPT (Tahiti: Society archipelago), TBI (Austral Islands), RAR (Cook Islands), and RKT (Gambier Islands). Two successive although overlapping sequences occurred in 2011. In 2013, only one was observed, its amplitude a little weaker and masked at RKT by strong microseismic noise. The frequency of the signal was perfectly monochromatic, at 0.0589 Hz (17.0 s), and was identical in the two cases.

in 2011 versus a single sequence in 2013. On the other hand, their similarities are rather striking. They include (i) similar spectra of relatively low-frequency volcanic tremor with a maximum energy between 0.5 and 4 Hz; (ii) correlation between the waves of short and long periods, the latter delayed from the former by about 10 min; and (iii) perfectly monochromatic Rayleigh waves of long period, with the same frequency of 0.0589 Hz (17.0 s) in 2011 and 2013. For a duration of 50 min, the frequency resolution would be $df = 3.3 \times 10^{-4}$ Hz ($df = 1/\text{duration}$). These resonant frequencies are thus very well defined and imply the resonance of the same structure during two separate events 2 years apart.

3. Discussion

3.1. Constraints on the Nature of Seismic Sources

Although the volcanic origin of the events described above is obvious given their locations, they clearly differ from all the volcano-seismic events previously recorded from the Society hot spot since monitoring began by the RSP half a century ago. Recordings preceding or associated with eruptions on the submarine flanks of Teahitia include harmonic tremor of high frequencies having a maximum energy between 5 and 7 Hz and spasmodic tremor of the same frequencies mixed with numerous small earthquakes. The *P* phases of the latter are associated with always larger *S* phases [*Talandier and Okal*, 1984, 1987; *Talandier*, 2004]. These volcanic tremors can be compared with those from Rocard recorded in 2011 and 2013 (Figure 2).

The spectra of the latter are rich in components of low frequencies presenting a maximum energy between 0.5 and 4 Hz that are mixed with some impulsive events of high frequencies, which are barely identifiable from the recordings of the stations located on Tahiti. In contrast to the low-frequency events, the Pb or Pn

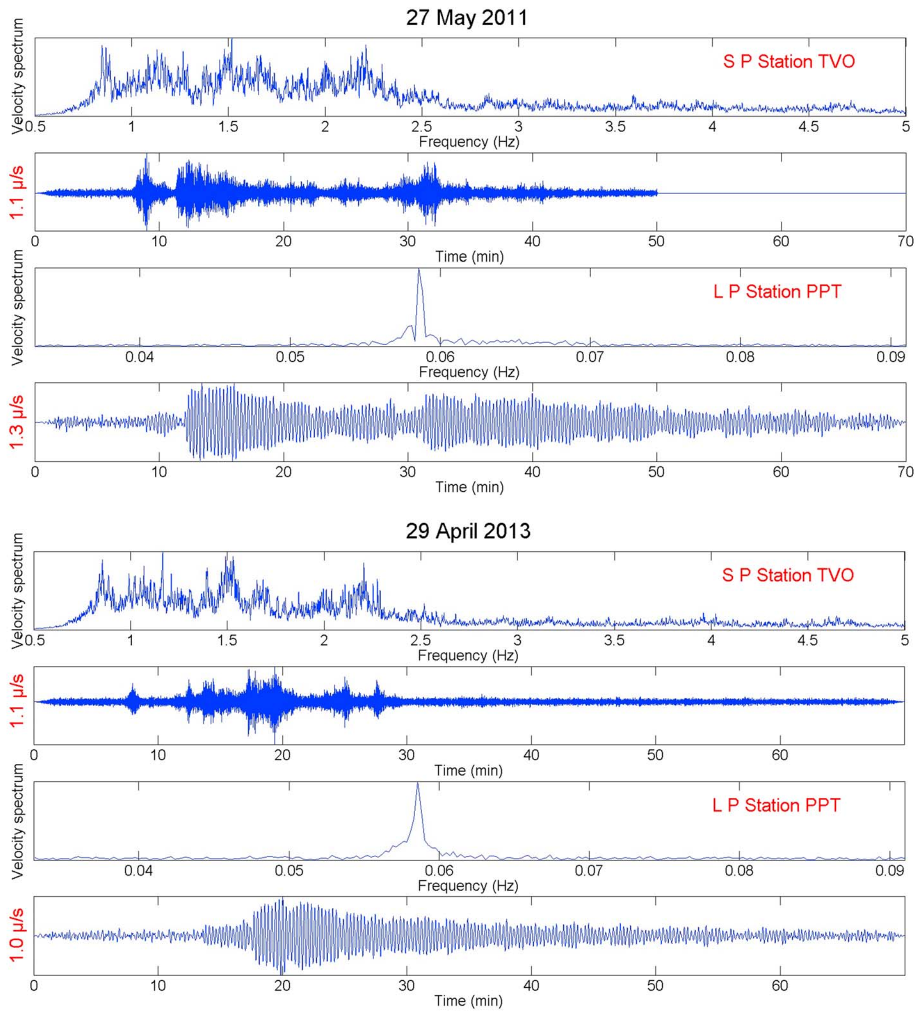

Figure 4. Comparison of the 2011 (top) and 2013 (bottom) seismic events: (a) frequency spectrum and (b) short-period events recorded at TVO (80 km away from the source); (c) frequency spectrum and (d) long-period events recorded at PPT (110 km away from the source). In 2011 and 2013, the short-period events preceded the long-period Rayleigh wave train by about 10 min. Short and long-period records are referenced at the same time.

waves of the rare impulsive events have a relatively narrow spectrum between 3 and 6 Hz and are almost devoid of associated *S* waves. This last feature allows us to discriminate them from the small earthquakes associated with the spasmodic volcanic tremor from seismic crises at Teahitia, where *S* waves dominate the signals.

The seismic signals associated with the eruptive processes were observed during the strong volcano-seismic crises of Mehetia in 1981 and Teahitia in 1982, 1983, 1984, and 1985, culminating with the emplacement of lavas on the submarine sides of these volcanoes [*Talandier and Okal*, 1984, 1987; *Talandier*, 2004]. The seismicity directly connected to an eruption was mainly due to the magma transfer from the upper mantle to the crust and the volcanic edifices, like in the Hawaiian Islands. Swarms of numerous small earthquakes were accompanied by volcanic tremor during the various stages of this process. At the end of the crisis, the lava effusions were generally followed by a more diffuse seismicity consisting of a small number of shallow earthquakes of stronger magnitude that were related to isostatic adjustments within the volcanic edifice. The high-frequency tremor is generally attributed to the resonance of cracks opened in the rock under the influence of magmatic pressure [*Aki et al*., 1977; *Aki and Koyanagi*, 1981]. In Teahitia, this high-frequency tremor is characterized by variable, and usually fluctuating, intensity, and they are frequently concomitant with a fast succession of earthquakes which mark the fracturing of a solid environment as suggested by the strong intensity of the associated *S* waves. The absence of these *S* waves from the impulsive events of 2011 and

2013 at Rocard suggests a sudden pressure drop in a viscous environment rather than stress drops characterizing earthquakes in swarms and those that accompany the spasmodic tremor at high frequency.

The most fascinating feature of the 2011 and 2013 seismic activities at Rocard volcano is certainly the long-period Rayleigh wave train that are perfectly monochromatic, intense, and of long duration. Obviously, contrary to the frequency-dispersed Rayleigh wave trains, the duration of these monochromatic wave trains corresponds to that in the source and the fluctuations in amplitude result from the variations of its intensity. A number of authors have previously reported long-period oscillations of volcanic origin, but these generally occur in the shape of impulses (Stromboli: *Chouet et al*. [2003], Merapi: *Hidayat et al*. [2002], Miyake: *Kumagai et al*. [2001], Kilauea: *Kumagai et al*. [2005], and Popocatepelt: *Chouet et al*. [2005]), as wave trains of short duration [*Kawakatsu et al*., 1994], or as volcanic tremor, (Usu: *Yamamoto et al*. [2002]). In most cases they consist of a train of quickly dampening waves resulting from the resonance of a conduit by an impulsive trigger (Erebus: *Rowe et al*. [1998], *Aster et al*. [2003], Galeras: *Gil Cruz and Chouet* [1997], Iwate: *Nishimura et al*. [2000], and Hachijo: *Kumagai and Obara* [2003], *Kumagai* [2006], *Kumagai et al*. [2003], and *Kumagai and Chouet* [2000]). Numerous studies have described long-period signals associated with the volcanism of the Hawaiian Islands [*Chouet*, 1985, 1981; *Chouet et al*., 2010; *Chouet and Dawson*, 2011]. However, none of these seismic events were characterized by perfectly monochromatic oscillations maintained for almost 1 h such as those that occurred during the 2011 and 2013 events at Rocard volcano.

3.2. Comparison of the 2011 and 2013 Rocard Events With the Activity Observed at Kilauea, Hawaii

For the interpretation of the 2011 and 2013 Rocard events we will refer mainly to the analyses by *Chouet et al*. [2010] of the explosive degassings of Kilauea and of *Chouet and Dawson* [2011]. Although these studies describe the seismicity of an oceanic hot spot similar to the Society hot spot, there are major differences between the Rocard and Kilauea volcanic edifices regarding their shape and size, the frequency and importance of their eruptive activities, and the configuration of the monitoring networks.

While the summit of Kilauea is at 1047 m, the Rocard summit is located 3200 m below sea level and therefore is subject to external pressures of 320 bars up to 400 bars at its base at −4000 m. Kilauea is almost continuously active and emits large volumes of lava (more than 3.5 km^3 since the start of the Pu'u 'O'o eruption in January 1983). In contrast, the 2011 and 2013 seismic events mark the first probable eruptions that can be attributed to Rocard volcano since the setup of the RSP network in 1963, with the possible exception of a very weak seismic episode in 1972 [*Talandier and Kuster*, 1976]. A dense network of seismic stations on Kilauea makes sophisticated analysis of individual events possible. In contrast, Rocard is located 70 to 100 km away from the closest stations of the RSP, and thus, it is difficult to record events with weak energy (Figure 1).

In the short-period frequency domain, explosive degassings of Kilauea are recorded as distinct individual events occurring randomly during an ongoing eruption, whereas the activity of Rocard occurred during two episodes 2 years apart, each of them involving a single series of events of fluctuating intensity during less than 1 h (Figure 4).

In the long-period frequency domain, individual explosive degassings at the Kilauea infer an oscillation of very low frequencies (0.04 to 0.1 Hz) damped within 10 periods. These oscillations of variable frequencies imply resonators of different dimensions. At Rocard, a single perfectly monochromatic wave train propagated Rayleigh waves over long distances for nearly an hour. In contrast to the Kilauea cases, these oscillations are maintained during the crises, which involved some 180 and 140 individual oscillations, respectively.

Finally, the signals of short and long periods recorded by the stations located on Kilauea were not received by the station POHA, only 47 km distant. In contrast, the Pn waves of volcanic tremor stemming from Rocard were recorded up to 100 km away, those of the impulsive events up to 200 km away, and the Rayleigh wave trains of long period up to least 1600 km away. These observations suggest that the events at Rocard were much more energetic than the individual explosive degassing events at Kilauea.

3.3. Possible Origins of the Rocard Seismic Signals

Chouet et al. [2010] showed that the complex seismic signals associated with long-period events observed in the Hawaiian Islands are related to the eruptive activity of Kilauea, characterized by bursts of explosive degassing. Such a mechanism cannot obviously be considered in the case of Rocard submarine volcano, because it is located at depths of 3200–4000 m below sea level. The 2011 and 2013 signals from Rocard also have little in common with those of known seismovolcanic crises of active submarine Polynesian hot spot volcanoes, e.g., Mehetia in 1981; Teahitia in 1982, 1983, 1984, and 1985 (Society hot spot); and Pitcairn hot spot in 2001–2002 [*Hyvernaud et al*., 2003], which were attributed to eruptive processes involving progressive vertical migration of the seismicity and submarine effusion of lava on the flanks of the volcanoes. In addition, the durations of Rocard seismic activity (less than 1 h) are much shorter than the crises marking the progressive rise of basaltic magmas in dykes and channels below and within oceanic shield volcanoes. The latter last commonly for several days, weeks, or even months to years as exemplified by eruptive processes at Kilauea, Hawaii [*Chouet et al*., 2010; *Chouet and Dawson*, 2011], Piton de la Fournaise, La Réunion [*Peltier et al*., 2009; *Roult et al*., 2012], the 2011–2012 El Hierro eruption, the Canaries [*Klügel et al*., 2015], and the 2014–2015 Bárðarbunga eruption, Iceland [*Sigmundsson et al*., 2015].

In contrast, the temporal scale of the Rocard signals is roughly consistent with that of the shortest historical eruptions recorded on emerged volcanoes [*Simkin and Siebert*, 1994], the lower limit for a major eruption being the draining in less than 1 h of the Nyiragongo lava lake in 1977. Many eruptions at Piton de la Fournaise last 1 day or less, and they are attributed to the sudden draining of a meter-wide shallow dyke located within the volcanic edifice, above sea level [*Peltier et al*., 2009; *Roult et al*., 2012; *Michon et al*., 2013]. For instance, the 21 June 2014 eruption within Enclos Fouqué, on the east-southeast side of the central Piton de la Fournaise cone, lasted ~20 h and was preceded by a 74 min long seismic crisis [*Global Volcanism Program*, 2015a, 2015b]. The initial stage of the Bárðarbunga eruption lasted only 4 h on 29 August 2014 and was followed by a 2 day pause before resuming from 31 August 2014 to 27 February 2015 [*Sigmundsson et al*., 2015]. A length of slightly less than 1 h is thus not unreasonable to envision at Rocard for the continuous outpouring of lava from a meter-wide shallow dyke.

With the magmatic pressure impeded by a hydrostatic pressure of ~400 bars, we interprete that the entire conduit could vibrate continuously during the ejection of lava as a result of the excitation of its resonance in the form of an intense long-period Rayleigh wave train. The delay of about 10 min of this train with respect to the preceding short-period volcanic tremor could correspond to the time necessary for the filling of the magmatic conduit before the beginning of lava outpouring. The energy required to overcome the hydrostatic pressure could explain the intensity of the Rayleigh waves recorded at great distances away from the edifice. The most intriguing, and still rather enigmatic, feature of the Rocard activity is the almost exact duplication of the phenomena after a 2 year pause. In oceanic shield volcanoes, most eruptions occur within radial rift zones that channel basaltic magmas within fractures, generating dyke arrays. The average widths of these individual dykes range from 40 cm to 1 m [*Annen et al*., 2001]. Such rift zones are evident on Society hot spot submarine volcanoes [*Binard et al*., 1991]. The 2011 and 2013 seismic events could be due to magma outpouring from two dykes of similar width possibly belonging to a single array within a radial rift zone. However, the perfect coincidence of resonance frequencies (0.0589 Hz) would require that these two dykes had exactly the same geometry, which is rather unlikely. Alternatively, the same conduit (either a dyke or a vertical cylindrical conduit) might have been used twice by ascending magmas separated by a 2 year pause. However, it is usually envisioned that at the end of a basaltic eruption, the feeding conduit is sealed by the solidification of lava and that the next magma batch will ascend through another (possibly neighboring) fracture, generating a new dyke. Nevertheless, downward fluxes of magma in dykes have been documented in oceanic shield volcanoes using the Anisotropy of Magnetic Susceptibility (AMS) technique, e.g., in Hawaii [*Knight and Walker*, 1988], the Azores [*Moreira et al*., 2015], and in the Canaries, especially in the Tenerife rift zones [*Delcamp et al*., 2015 and references therein]. They have been interpreted as primary downward fluxes resulting either from lava drainback after emplacement or from magma emplacement above the level of neutral buoyancy [*Delcamp et al*., 2015 and references therein]. It is thus not unreasonable to postulate that such magma withdraw at the end of the 2011 event may have left an open fracture that was later reused for the ascent of the 2013 magmas. We admit that this scenario may look rather unusual, but the similarities between the 2011 and 2013 Rocard signals are also strikingly unusual.

4. Conclusions

The unusual long-period seismic events at Rocard seamount in 2011 and 2013 cannot be explained by previously observed eruptive processes such as those observed in the Hawaiian Islands [*Dawson et al*., 2010] or associated with other Society hot spot submarine eruptions, e.g., Mehetia in 1981 and Teahitia in 1982, 1983, 1984, and 1985 [*Talandier and Okal*, 1984, 1987; *Cheminée et al.*, 1989; *Talandier*, 2004]. We tentatively interpret that the observed intense Rayleigh wave trains that were strikingly monochromatic and lasted nearly an hour could be a consequence of the strong hydrostatic pressure acting on the rift zones of the submarine shield volcano. With the magmatic pressure impeded by this hydrostatic pressure, the transit of lava might have excited the resonance of a shallow opened conduit, likely a meter-wide dyke or a vertical cylinder. The delay of about 10 min between the signals of short (Pb-Pn waves) and long periods (Rayleigh waves) could correspond to the time lapse necessary for the filling of the conduit before the beginning of lava outpouring. The most intriguing, and still rather enigmatic, feature of Rocard activity is the almost exact duplication of the phenomena, with an identical frequency of the oscillation of Rayleigh waves (0.0589 Hz) after a 2 year pause. We tentatively propose that the 2011 and 2013 events were due to shallow magma outpouring from the same conduit in both events, despite the questionable geological plausibility of such a process.

Acknowledgments
We dedicate this study to the memory of Yves Rocard who created the Laboratoire de Détection et Géophysique and supported the development of the Tahiti laboratory and its seismic network. We thank Bernard Chouet and Emile Okal for fruitful discussions. Many thanks to William Chadwick for his very helpful and detailed review. Waveforms are available at the Laboratoire de Géophysique (contact hyvernaud@labogeo.pf).

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