**Testing potential trigger mechanisms for seismicity in Sarria-Triacastela-Becerreá (Lugo seismic sequences) NW Iberian Peninsula, Spain**

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**Abstract**

The unprecedented and long-lasting seismicity of Triacastela (over 25 years) attracted the interest of the research community on this stable continental region (SCR), particularly after the anomalously high 5.1 and 4.9 Mw earthquakes compared to the regional standards. The high rainfall and tide rates of this region compared to the rest of the Iberian Peninsula, in addition to the recognized existence of thermal springs and crustal fluids, motivated us to test these sources of hydroseismicity as a potential trigger mechanism for the observed seismicity in Triacastela. Based on network upgrades, we have gathered the seismic catalog in two periods for analysis (before and after year 2002). Before 2002, neither a diffusion-type earthquake migration nor any significant statistical correlation between the seismicity rates and rainfall or tides is found. After 2002, some clusters migrate to the south, suggesting the presence of fluid migration during earthquake swarms, but no diffusion-type migration is observed on longer time scales. Furthermore, we find correlation coefficients close to zero, indicating that rainfall and tides can be excluded as driving mechanisms. However, the seismic upward migration, a high b-value (1.2), and a low aftershock-productivity parameter ($\alpha$=0.9) observed in this period support the hypothesis of upward fluid

migration through fracture zones. The presence of Mantellic helium-3 along the seismogenic faults and the increase of geochemical precursors in the groundwater previously to 1995 and 1997 mainshocks further support deep fluids as a source for the observed induced fluid migration seismicity in Triacastela.

**Article Highlights:**

- Swarms, upward migration, high b-value, and low α-value after 2002 suggest a fluid influence on Triacastela seismicity.
- Pearson correlations close to zero reject rainfall and tides as driving mechanisms of Triacastela seismicity.
- The mantellic helium-3 in groundwaters, the geochemical precursors rise and elements mobility over faults evoke deep fluids migration.

**KEYWORDS**

Induced seismicity, hydraulic diffusivity, fluid migration, deep fluids, rainfall, tide

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## 1. Introduction

The seismicity of Sarria-Triacastela-Becerreá (Lugo seismic sequences) is located in an intraplate region considered seismically stable (NW Iberian Peninsula), but in November 1995 started an unprecedented and clustered seismicity never registered before in this area. The anomalous long-term seismicity continues until now (over 25 years) and involves the 48% of the seismic moment release in the whole NW Iberian Peninsula (e.g., Martínez-Díaz et al. 2006; Martín-González et al. 2012; Crespo-Martín et al. 2021).

Combined with the long-lasting seismicity, the existence of the highest magnitudes compared to the regional standards (5.1 Mw) accomplish the updating of the Spanish Building Code (NCSE 2002).

Plate boundaries typically register the highest earthquake rates and magnitudes, whereas seismicity also occurs inside the tectonic plates, but its origin is poorly understood (Calais et al. 2016; Maystrenko et al. 2020). The unexpected intraplate earthquakes produced even more damage due to the unprepared seismic-resistant buildings (e.g., Gupta 1998; Hough et al. 2002). In intraplate regions, the slow lithospheric deformation provokes low tectonic stressing rates, and it is just where the hydromechanical coupling could trigger seismicity (Kraft et al. 2006; Hainzl et al. 2013, 2015). The pore-pressure increase related to fluid intrusion and migration directly affects the stability of faults, especially those whose crustal stress is close to the critical failure (Shapiro et al. 1997; Hainzl et al. 2015). A straightforward proof of the instability of faults in intraplate environments is provided by, e.g., fluid injections in wells (e.g., Zoback and Harjes 1997; Rothert and Shapiro 2003; Brudzinski and Kozlowska 2019) and reservoirs induced seismicity (e.g., Talwani 2000; Hainzl et al. 2015). The occurrence of earthquake swarms is also assumed to be linked to this hydromechanical coupling in highly fractured regions where minor stress changes can produce failures (Mogi 1963; Kraft et al. 2006). Studies point out less-known sources of hydroseismicity. For instance, rainfall can increase in crustal stresses by loading pore-pressure diffusion due to surficial water infiltration (Muço 1995, 1999; Hainzl et al. 2013, 2015). Furthermore, the main components responsible for oscillating stress in the solid Earth are earth tides and seismic waves (Iwata, 2012). Earth tides can produce enough pressure to trigger earthquakes in critically stressed regions (e.g., Klein 1976; Wilcock 2009; Hainzl et al. 2013; Ide et al. 2016). Several inquiries prove how fluids could ascend into the upper crust across the preexisting faults and trigger seismicity (e.g., Miller 2013; Tang et al. 2021).

The NW Iberian Peninsula is characterized by the highest rainfall rates in the Iberian Peninsula (AEMET 2021a), which leads us to hypothesize that it potentially influences the recorded seismicity. Likewise, seismicity might also be affected by the Earth tides, which have, at the Atlantic coast (NW Iberian Peninsula), the largest amplitudes within Spain (Puertos del Estado 2021a). Triacastela region is considered as a fractured region where triggering by deep fluids can be expected (Almeida 1982, Pérez et al. 1996, 2008; Marques et al. 2003; Crespo-Martín et al. 2021). Considering the mentioned forcing potential of rainfall, tides, and deep fluids in the NW Iberian region, we look upon to understand the fluid component's origin. In this region, the dominant hypothesis of the triggering seismicity is the deep fluids migration (Crespo-Martin et al. 2021), however in our research a more comprehensive analysis by investigating also possible correlations with rainfall and tides is done. As documented worldwide, the role of fluids in triggering seismicity can be induced by human activity such as impoundment of reservoirs, (e.g., Stabile et al. 2014; Telesca et al. 2015), exploitation of hydrocarbons (e.g., Vilarrasa et al. 2021). None of these human activities are located in the Triacastela area as hydrocarbon or gas injection (Hydrocarbon Technical File 2022), impoundment of reservoirs (Confederación Hidrográfica Miño-Sil 2022), mining (IGME 2022), or even nuclear test that are forbidden in Spain since 1996 (CTBTO 2022), therefore we can discard a human factor behind the studied seismicity. Besides, no gas storage neither geothermal stimulated system operates in the study region (Ministerio para la Transición Ecológica 2017). However, our statistical research and methods are also of interest to human induced seismicity studies due to it allows identification of triggered earthquakes and associate them with a likely vector of induced seismicity.

Fluid processes are often associated with pore-pressure diffusion. Thus, an unequivocal observation for the presence of fluids is the detection of diffusion-type seismicity migration. A basic technique is to estimate the distance r of the propagation pore-pressure front from the source, which depends on the hydraulic diffusivity D (Shapiro et al. 1997; Parotidis et al. 2003). Besides, potential correlations between earthquakes and rainfall or tides can be revealed by the Pearson correlation coefficient that measures the linear relationship between two independent databases (Student 1908; Kowalski 1972; Scipy 2008-2018).

The aim of this work is to test the influence of rainfall, tides, and deep fluids as potential triggers of the seismicity in the region of Triacastela. For this purpose, we employ different techniques for the declustered earthquake catalogue. We analyze seasonal and daily variations of the seismicity, which could indicate atmospheric effects, and analyze the correlation with rainfall data. We repeat the correlation analysis also for tidal data. Finally, we search for migration patterns indicating pore-pressure diffusion. Based on our findings, we draw conclusions concerning the role of deep fluids.

**2. Geological and seismological framework**

Far away from the active plate boundaries of Eurasian and Nubian African Plate at the southern, the NW Iberian Peninsula is assigned an intraplate regime (Fig. 1) (e.g., Martín-González et al. 2012; López-Fernández et al. 2018; Crespo-Martín and Martín-González 2021). The NW Iberian Peninsula is the western termination of the Alpine Pyrenean-Cantabrian Orogen, whose main Alpine structures follow E-W thrust and NE-SW and WNW-ESE strike-slip faults (Martín-González 2009; Martín-González et al. 2021). After the Tortonian, the deformation migrated to the southern Iberian plate boundary and the region suffered the far-field effects of the southern border (Betics chains) (Galindo-Zaldivar et al. 1993; Martín-González et al. 2012, 2014). The seismogenic structures reactivated under an intraplate regime with the present stress field (oriented NW-SE), and caused the recent seismicity (e.g., Martín-Serrano et al. 1996; Martínez-Díaz et al. 2006; De Vicente et al. 2008; Martín-González and Heredia 2011; Martín-González et al. 2012). Since then, the spatial distribution of seismicity and focal mechanisms show that faults trending NE-SW to N-S and WNW-ESE seem to be responsible for most of the present seismicity (e.g., Martínez-Díaz et al. 2006; Martín-González et al. 2012; López-Fernández et al. 2018). The studied region is located in the Iberian Massif which is a Variscan basement of Precambrian and Paleozoic metamorphic sedimentary rocks mainly deformed during the Variscan Orogeny. The Variscan deformation phases are responsible for the main deformation in the area. The Alpine deformation in the study area is moderate. The Variscan Orogeny is responsible for the kilometric in amplitude recumbent folds with eastward vergence, the NNW–SSE brittle thrust towards the east, and the late extensional episodes with the same trend. Afterwards, before the middle Permian, in the western regions, different types of granitoids were intruded (González Lodeiro et al. 1982; Martínez Catalán et al. 1990; Barrera Morate et al. 1989). Since then, in the studied area there is no record of metamorphism, volcanism or massive tectonic events, and nowadays located more than 1 000 km from the active borders it is consider as a stable continental region (SCR) (Crespo-Martín and Martín-González 2021).

Despite considering a seismically stable region, the NW Iberian Peninsula shows a moderate intraplate seismicity with earthquakes up to 5.1, 4.9 and 4.6 Mw (e.g., Martínez-Díaz et al. 2006; López-Fernández et al. 2012; Martin-González et al. 2012). Several episodes of historical earthquakes occurred in this area, with EMS-98 (European Macroseismic Scale 1998) (Grünthal 1998) intensities between IV and X, which reflects the seismogenic potential of the region (Crespo-Martín et al. 2018). The densest distribution of events is located in Sarria-Triacastela-Becerreá (henceforth Triacastela,) which releases 48% of the seismic moment in the NW Iberian Peninsula in the last decades (Fig. 1). The highest magnitudes are registered in Triacastela: 5.1 Mw and 4.9 Mw in May 1997, and a doublet of 4.6 Mw in November and December 1995. The seismicity of Triacastela started in 1995 and continues steadily 25 years later in the same location (Martínez-Díaz et al. 2006; López-Fernández et al. 2012; Crespo-Martín et al. 2021). A mix of seismic mechanisms has been proposed as the origin of the seismicity: starting with tectonic seismicity followed by aseismic forcing (Crespo-Martín et al. 2021). From 1995 to 1998/99, the sudden fluctuations in background seismicity, the occurrence of deeper events, northeast migration, and the comparison to near tectonic seismicity indicate that the mainshocks triggered their aftershocks and facilitated or initiated aseismic processes. From 1999 onwards, even decreasing, the background activity and shallower events took control of the Triacastela seismicity. This result, along with the dominance of swarm-like behavior and high b-values, reflects aseismic forcing and a possible fluid migration as a driven mechanism of the seismicity after 1998/99 (Crespo-Martín et al. 2021).

POSITION OF FIGURE 1.

## 3. Databases: earthquakes, rainfall, and tides

### 3.1 Earthquake database

The earthquake catalogue used in our analysis is free access provided by the Spanish National Earthquake Catalogue that is compiled by *Instituto Geográfico Nacional* (IGN), the agency responsible for the alarm, Seismic Network, and earthquake catalogue (IGN 2021 - https://www.ign.es/web/ign/portal/sis-catalogo-terremotos). This database includes information from Latitude 42º 36'0" N, 43º17'60" N, and Longitude 7º 27´36" W, 7º 0'0" W. Our analysis period starts in 1988 and finishes on September 30$^{th}$, 2018, when 1475 earthquakes have been recorded (Fig. 1).

### 3.2 Rainfall database

We have compiled information about rainfall provided by the *Agencia Estatal de Meteorología* (AEMET). This institution is the Spanish Ministerial agency to provide weather forecasts and past meteorological data in open access service (AEMET 2021a - https://opendata.aemet.es/centrodedescargas/inicio). AEMET runs a dense weather network; in the Galicia region, 53 stations operate. For our analysis, we select the data of the meteorological station of Lugo Airport (43º6'41''N, 7º27''27''W) because of its proximity and the long-term database compared to the other nearby stations (Figs. 1 and 2a). The Lugo station provides information on daily rainfall (mm) from 1995 to 2018 (Fig. 2; AEMET 2021c - http://www.aemet.es/es/eltiempo/observacion/ultimosdatos?k=gal&l=1505).

POSITION OF FIGURE 2.

**3.3 Tidal database**

The tidal database is provided by the agency *Puertos del Estado,* belonging to the Ministry of Transport, Mobility, and Urban Agenda (Puertos del Estado 2021a - http://www.puertos.es/es-es/oceanografia/Paginas/portus.aspx). The long-term database information (from 1993 to 2018) and the spatial proximity guide us to select the tide gauge of Vigo2 (42º14'24''N, 8º43'48''O) (marked in Fig. 1). The Ministry provides several daily parameters to describe tides, taken the tidal level reference of Zero REDMAR (Cero of each tide gauge of *Red de Mareógrafos de Puertos del Estado* - REDMAR): (1) Mean Level (cm): value resulting to remove the diurnal and semi-diurnal component of the daily tide data and performing a subsequent filtering of 119 points centered on noon; (2) High Tide (cm): maximum level of the daily astronomical high tide; (3) Low Tide (cm): minimum level of the daily astronomical low tide; (4) Max. Tide Range (cm): the maximum vertical difference between high tide and low tide; and (5) Min. Tide Range (cm): the minimum vertical difference between high tide and low tide (Fig. 3) (Puertos del Estado 2021a - http://www.puertos.es/es-es/oceanografia/Paginas/portus.aspx); Puertos del Estado 2021b - https://bancodatos.puertos.es/BD/informes/globales/GLOB_2_3_3221.pdf).

POSITION OF FIGURE 3.

Some data gaps exist for the Vigo2 station. Thus, we applied linear interpolation using information of the nearby Villagarcia2 station (42º36'0''N, 8º46'12''O). For this purpose, we first performed a linear regression between each tidal parameter of Vigo2 and Villagarcia2 stations and then inserted missing data accordingly (Table 1).

The two stations show a high correlation (R value>0.95), justifying our procedure for filling the data gaps.

If data of both stations missed simultaneously, we simply apply a linear interpolation between prior and following days according to

$$Y = Y1 + \frac{Y2-Y1}{t2-t1} \cdot (t - t1)(1)$$

where Y is the unknown value at time t. Y1 and Y2 are the prior and following values at time t1 and t2, respectively. Using this procedure, the tidal database is completed for the period from 1993 to 2018. For illustration, Fig. 3 shows the resulting tidal parameters for the year 2002.

**4. Methods**

**4.1 Preparing the earthquake catalogue**

4.1.1. Homogeneous and completed catalogue

Before analyzing the seismicity, it is crucial to homogenize and bear in mind the upgrades of the earthquake catalogue (Naylor et al. 2010). Especially our study covers a long-time period of seismicity, from 1988 to 2018 (Naylor et al. 2010). Before 2002, four analog short-period seismic stations (vertical component) operated: STS, ERUA, EMON and EZAM. The first three stations closed in 2010, 2009 and 2002 respectively, while EZAM station was replaced to a broadband network in 2012. From ending 2001 to 2002, six new digital broadband stations (three-component) started to operate in the region: ELOB, EARI, ECAL, EPON, EMAZ and EINC. The close down of EINC in 2005 and the opening in 2010 of a new broadband station (EAGO) complete the actual framework of IGN seismic stations in the NW Iberian Peninsula (Fig. 1) (González 2016; Crespo-Martín et al. 2021).

Caused by the improvements of the seismic network, a relevant improvement of the event detection with a corresponding drop of the completeness magnitude Mc occurred in 2002, which entails the distinction of two periods. Before 2002 (1988-2002), the seismicity is dominated by unprecedented clustered seismicity that started in November 1995. A completeness magnitude of 3.2 is calculated for this period, and 206 events are registered (Mc≥3.2). From 2002 until its end on 30$^{th}$ September 2018, the catalogue contains smaller magnitude events due to an upgraded seismic network. In this latter period, the Mc drops to 1.8 and more events are registered (722 events Mc≥1.8) (Crespo-Martín et al. 2021). Here, Mc is estimated with the MAXC technique using the MATLAB code ZMAP (Wiemer 2001; Mignan and Woessner 2012).

During the evolution of the seismic monitoring network, different magnitude scales have been used, such as MbLg (M-MS), mb (V-C), mbLg (L), and Mw (González 2016). A systematic comparison of the seismicity over time requires homogenization of the earthquake size parameter. This homogenization into Moment Magnitude (Mw) was done with the updated equations of the Spanish Seismic Hazard Map (IGN-UPM 2013; Cabañas et al. 2015).

#### 4.1.2. Relocated catalogue

To analyze the diffusion-type migration patterns and to analyze the depth evolution, we relocated earthquakes of Triacastela. This method was previously applied in Triacastela in the study of Crespo-Martín et al. (2021). Taken the homogenized and completed catalogue, we used the technique of double-difference DD relocation according to the HypoDD algorithm, which is written in Fortran language (Waldhauser and Ellsworth 2000). The conjugate gradient method (LSQR) is applied to solve the DD equations (Paige and Saunders 1982; Waldhauser and Ellsworth 2000; Waldhauser 2001). As it is suggested by Waldhauser (2001), to assess the accuracy of the relocation errors, we divided the catalogue into subsets of clusters and recalculated the hypocentre relocation with the singular value decomposition technique. This study used the temporal P and S-wave arrivals and seismometers x, y position as input data.

Considering 1475 earthquakes registered from January 1988 to September 2018, we relocated 975 earthquakes. Maximum standard deviations are ± 0.46 km (Ex), ± 0.66 km (Ey), and ±0.70 km (Ez).

#### 4.1.3. Declustered catalogue

After getting the homogenized and complete earthquake catalogue, we identified the externally triggered seismicity using the Epidemic-Type Aftershock Sequence (ETAS) model. According to their trigger, this technique classifies earthquakes either as so-called aftershocks triggered by preceding earthquakes or as so-called background events triggered by external sources (van Stiphout et al. 2012). Only the background events can be expected to be directly related to transient stresses induced by fluid intrusions, rainfall, and tides. The ETAS approach allows assigning a probability to each earthquake being a background event or an aftershock (Zhuang et al*.* 2002; Ogata and Zhuang 2006; van Stiphout et al*.* 2012). Specifically, we used the ETAS approach developed by Marsan et al. (2013), which accounts for time-dependent background rates (Hainzl and Ogata 2005; Hainzl et al. 2013).

Here, we use the declustered catalogue derived by Crespo-Martín et al. (2021) to study potential correlations with rainfall and tides.

**4.2. Pore-Pressure Diffusion**

When the crustal is in a critical stress state, slight perturbations can already trigger seismicity. Adjustments of effective normal stress and the fault's strength are caused by increased pore-pressure due to a fluid injection (Shapiro et al. 1997). The diffusion equation defines pore-pressure variations due to a fluid intrusion from a high-pressure source (e.g., Shapiro et al. 1997; Hainzl and Ogata 2005).

$$\frac{\partial}{\partial t}P = D\frac{\partial^2}{\partial x^2}P \qquad (2)$$

where D is the hydraulic diffusivity, which tends to be between 0.01 and 10 $m/s^2$ in the crust, P is the pore-pressure, and t is the elapsed time from the first contact of the pore-pressure source with the host rock (Shapiro et al. 1997).

We applied the method developed by Shapiro et al. (1997) to solve this equation, which considers a point source in a homogeneous isotropic saturated poroelastic medium. They estimated the distance r of the propagation pore-pressure front from the source as

$$r = \sqrt{4\pi D t} \qquad (3)$$

This equation represents a parabolic curve in the r-t plot that is a hallmark for seismic swarms triggered by pore-pressure diffusion. The earthquakes are expected to mainly occupy the space below this parabolic envelope (Shapiro et al. 1997; Parotidis et al. 2003; Hainzl and Ogata 2005).

The variable r in the r-t plot is the distance of the event's hypocenter (with index i) to the reference point ($X_0$, $Y_0$, $Z_0$).

$$r = \sqrt{(X_0 - X_{i-th})^2 + (Y_0 - Y_{i-th})^2 + (Z_0 - Z_{i-th})^2} \text{ (4)}$$

where X refers to the N-S direction (latitude), Y represents the E-W direction (longitude), and Z is depth, all in units of km. Seeking migration in a specific coordinate, we also estimate the X =($X_{i\text{-}th}$- $X_0$), the Y=($Y_{i\text{-}th}$- $Y_0$), and the Z=($Z_{i\text{-}th}$ - $Z_0$) migration, where the last one is relevant for rainfall triggering.

In this study, we based the measure of hypocentral distance on the first event of each period (29/11/1995 before and 01/02/2002 after 2002) (Fig. 6).

**4.3. Pearson's correlation**

Pearson correlation coefficient is a well-known statistical tool to measure the linear relationship between two independent databases, which follow a normal distribution. The correlation analysis provides two outcomes: the Pearson's correlation coefficient (C-Pearson) and the p-value. C-Pearson achieves values from +1 (positive correlation) to -1 (negative correlation), with 0 implying no correlation and is calculated by

$$r = \frac{\sum(x-m_x)(y-m_y)}{\sqrt{\sum(x-m_x)^2 \sum(y-m_y)}^2} \qquad (5)$$

where $m_x$ and $m_y$ are the means of x and y, respectively. Under the assumption that x and y come from independent normal distributions, the probability density function of r is described by

$$f(r) = \frac{(1-r^2)^{n/2-2}}{B\left(\frac{1}{2},\frac{n}{2}-1\right)} \qquad (6)$$

where n is the sample number and B is the beta function (Student 1908; Kowalski 1972, Scipy 2008-2018). For a given coefficient r, the p-value then represents the probability that the null hypothesis (no correlation) is correct. Here, the p-value is two-sided. A small p-value (<0.05) indicates that the null hypothesis is incorrect, and the observed correlation is statistically significant (CORSSA 2021. http://www.corssa.org/en/glossary/).

For the application of the Pearson correlation, the two considered time series have to be firstly binned in the same way. For the earthquake database, we have to distinguish the two temporal periods: 1995-2002 and 2002-2018. The tidal database gives information about several tide parameters (Max. Tides, Min. Tides, Mean, High, Low tides). The Pearson correlation is calculated for these tidal parameters individually. For both tides and rainfall, we have tested bin sizes of 1, 10, 20, and 30 days. Furthermore, a potential delay can be considered, which can account for a delayed triggering mechanism. We only consider instantaneous triggering by tides (no shifts) because the response is expected to be immediate due to tidal stresses act simultaneously at all depths. In contrast, a time shift is more reasonable for rainfall because the water needs time to permeate into the ground and potentially trigger earthquakes. Therefore, we analyzed time shifts between 1 to 100 days between rainfall and seismicity.

**4.4. Gutenberg-Richter Law and time-dependence background ETAS model**

The Gutenberg-Richter Law proposes an approach of the seismic dataset (Vere-Jones, 2010). It relates the magnitude and the number of events of at least this magnitude, and it is defined by the Equation 7 (Gutenberg and Richter 1956).

$$log_{10}N(\geq m) = a - bm \qquad (7)$$

Where $N(\geq m)$ is the cumulative number of events with magnitudes greater than *m*. The *a-value* is a constant and *b-value* denotes the seismic properties linked to the stress regime (Gutenberg and Richter 1956; Vere-Jones 2010). In this research, the Gutenberg-Richter parameters are estimated applying the Maximum Likelihood method (Aki 1965), and the Magnitude of Completeness (Mc) with the Maximum Curvature Estimation (Wiemer and Wyss 2000).

The *Epidemic Type Aftershock Sequence* (ETAS) model, proposed by Ogata (1988) considers two contributions of earthquake rate: aseismic forcing or background contribution (μ), and earthquake-earthquake interaction term (ν) (Eq. 8) (e.g., Ogata 1988; Zhuang et al. 2012).

$$\lambda(t) = \mu + \mathrm{v} = \mu + K\sum_{t_i<t}\frac{e^{\alpha(m_i-M_c)}}{(t-t_i+c)^{p'}} \qquad (8)$$

Where the interaction earthquake term (ν) sums over all earthquakes i (with occurrence time ti and magnitude mi) occurred before time t. The α-value (magnitude-1) defines the efficiency on an event of a given magnitude in generating aftershock. K-value denotes the productivity of an event with threshold magnitude Mc, while c (unit of time) and p´-value determine the decay rate. The Marsan et al. (2013) approach applied in this research follows the Equation 8 and suggests a time-varying forcing rate, μ(t). It is used an iterative algorithm to estimate ETAS parameters and time-dependent background by using the maximum likelihood estimate (Marsan et al. 2013).

## 5. Results

### 5.1. Seasonal variations

At first, we analyzed potential seasonal and monthly variations of the declustered earthquake activity. The aftershocks were removed to avoid any bias due to earthquake-earthquake interactions. The percentages of earthquakes in the different seasons are provided in Table 2, while the monthly results are shown in Fig. 4. The results are compared to randomized catalogues. For that purpose, we consider a stationary Poissonian process with the mean rate of the declustered catalogue. The corresponding standard deviation of the uncertainties is simply given by the square-root of the rate. In Fig. 4, the result for a Poisson process is indicated by the horizontal red lines with error bars. Before 2002, the activity has a significant peak in May for both complete and background events. This peak cannot be explained by a random fluctuation. However, a detailed analysis shows that this peak is mainly related to the activity in May 1997 and does not represent a general seasonal behavior. After 2002, no significant seasonal variation can be observed for the declustered activity (Fig. 4).

POSITION OF TABLE 2.

POSITION OF FIGURE 4.

### 5.2 Tidal correlation

To determinate the conditions leading to the highest C-Pearson value, we consider the absolute value of the correlation coefficient. Non-significant correlations indicated by p-values greater than 0.05 are found for Max. Tidal, Min. Tidal, High Level, and Low Level. Statistically significant values are only found for the Mean Level. However, in this case, the Pearson correlation is close to zero for daily binning ($\Delta t = 1$): $C=0.05$ with $p=0.03$ (1995-2002) and $C=-0.03$ with $p=0.02$ (2002-2017), indicating no relevant correlation between earthquakes and tides for both periods. Some correlation is displayed for a binning of 20 days ($\Delta t = 20$) from 1995-2002. In this case, a significant correlation coefficient of 0.20 with $p=0.03$ is found. However, for slightly different binning ($\Delta t$ =10 or 30 days), no significant correlations are observed ($p>0.05$), suggesting that the result for $\Delta t$ =20 days occurs by chance (Table 3; Fig. 5).

POSITION OF FIGURE 5.

POSITION OF TABLE 3.

**5.3 Diffusion-type migration**

The most certain indication for fluid-triggered seismicity is the detection of migration patterns in agreement with pore-pressure diffusion (Parotidis et al. 2003). To that end, we firstly measured hypocentral distances relative to the first event of each period (29/11/1995 before and 01/02/2002 after 2002) as a function of the occurrence times relative to those events (29/11/1995 before and 01/02/2002 after 2002) (Fig. 6).

POSITION OF FIGURE 6.

The fit of the theoretical diffusion curve (Eq. 3) of fluid diffusion to the entire period is poor, with best estimates of the hydraulic diffusivity of D=0.5 $m/s^2$ and 0.01$m/s^2$, before and after 2002, respectively (Fig. 6). Some of the swarms after 2002 seem to indicate migration (marked in boxes in Fig. 6b). 1.- April2004-Aug2004 (around day 900), 2.- Sep2005-Aug2006 (1300-1800 days), 3.- Dec2007-Feb2008 (2000-2200 days), 4.- Febr2010-Apr2010 (2900-3000 days), and 5.- Jan2013 (around day 4000) (Fig. 6b, and Table 6). However, we find that also these swarms do not clearly follow Eq. 3 and thus cannot be with certainty attributed to pore pressure diffusion.

**5.4. Rainfall correlation**

We repeated the correlation analysis for the measured rainfall at station Lugo. In particular, we tested temporal bins of 1, 10, 20, and 30 days. In all cases, we calculated the Pearson's correlation coefficient of zero-time shift. However, in the case of daily bins, we also calculated the correlation coefficients for time shifts between 1 and 100 days.

All results are summarized in Tabs. 4 and 5. We observe only for the daily data after 2002 a significant correlation with p=0.01 smaller than 0.05, while all other values are not significant. However, the correlation coefficient is almost neglectable in the former case with a value of -0.03. Thus, we also consider this result insignificant. Considering potential delays does also not lead to significant correlations. The highest correlation coefficient (-0.03) is obtained without delay time with p=0.01, suggesting no significant correlation.

Although the overall comparison does not indicate a systematic correlation between seismicity and rainfall, the daily plot of rainfall and earthquake rates reflect a possible correlation of both variables in some clusters associated with the highest precipitation peaks: 74, 73, and 68 mm on 9$^{th}$ June 2010, 18$^{th}$ September 1999, and 18$^{th}$ January 2013, respectively (Fig. 7).

POSITION OF FIGURE 7.

Before 2002, the clustered seismicity does not seem to coincide with the highest rainfall events (Fig. 7a), while some clusters appear to correlate in the second period (Fig. 7b). This result might be related to our observation that after 2002 some clusters show migration patterns (Fig. 6). Qualitatively, Fig. 7 shows that some clustered seismicity could be influenced by rainfall, such as January 2013 (4000 days relative to 12/01/2002), which also shows migration pattern in Fig. 6b. To quantify and assure this influence, we calculated the cross-correlation coefficients of rainfall and earthquake rate for the most prominent clusters. However, similar to the case of the long periods, the resulting high p-values for the clusters (varying between 0.1 and 0.9) do not indicate any significant correlation with rainfall (Table 6).

POSITION OF TABLE 4.

POSITION OF TABLE 5.

POSITION OF TABLE 6.

**5.6 Deep fluids**

Employing the relocated catalogue, an upwards migration of the seismicity is found over the years. The mainshocks of 1995 and 1997 are deeper than the seismicity in the following years when the earthquakes decrease in-depth and migrate towards the surface, especially after 1999 (Fig. 8).

POSITION OF FIGURE 8.

POSITION OF TABLE 7.

The fast decay of the aftershocks rates reflected by Omori's p'-values between 1.1 and 1.4 are typical to intraplate regions (1.1 to 1.7) (Utsu 1969; Ogata 1989, 1992; Utsu et al. 1995). From 2002 onwards, the Mogi Classification (1C and 2C) and an aftershock productivity parameter $\alpha$ of 0.9 (<1) reflect a swarm-like behavior. These characteristics contrast with the 1995 and 1997 mainshock-aftershock sequences (1A and 2A) with $\alpha$ >1 (from 1.03 to 1.57). The b-value higher than one (1.2 ± 0.04) compared to the close to one b-values of the mainshock-aftershock sequences (1.0 and 1.1) also indicates a swarm activity and could suggest a non-tectonic origin such as fluid migration through a fractured heterogeneous crust (Table 7) (e.g., Mogi 1962; Scholz 1968; Hainzl and Fischer 2002; Hainzl et al. 2013).

**6. Discussion**

The highest precipitation rate and tidal fluctuation compared to the Iberian Peninsula motivated us to check the likelihood of rainfall and tides as a trigger mechanism of Triacastela seismicity. The sudden fluctuations of the background rate switching towards a stationary process and the increase of the background contribution to 55% suggest that the largest mainshocks of 1995 and 1997 initiated or facilitated aseismic forcing that later continues, especially after 1998/99 (Crespo-Martín et al., 2021). Based on the shift of the seismic network, we defined two periods for our analysis of potential correlations with rainfall or tides.

Before 2002, monthly and seasonal fluctuations are shown in May-June and December, which could be linked to the main sequences of May 1997 and December 1995 (Table 2 and Fig. 4a). In this period, pore-pressure diffusion provides neither a good fit on a long-timescale nor for short-term clustering (Fig. 6a). Furthermore, no significant correlation between seismicity and rainfall could be found, and almost all correlations with tidal variations are insignificant. Statistically significant values are only found for some particular binning for the Mean Level tides. However, the correlation of the Mean Level is insignificant for other bin sizes, and the correlations for the other four tidal parameters (maximum and minimum tidal ranges and levels) are insignificant in all cases. These results exclude tides and rainfall as responsible trigger mechanism for the seismicity before 2002 (Tables 3, 4, and 5).

From 2002 onwards, the seismicity is characterized by a high b-value (1.2), a low α-value (0.9), and the dominance of swarm activity (Table 6). In literature, these observations are linked to hydromechanical coupling suggesting an influence of fluids (Kraft et al., 2006). However, the seismicity does not show a clear long-term diffusion-type migration pattern. Specifically, the earthquake locations cannot be fitted by Eq. (3), indicating that no systematic migration took place over long time. On short time-scales, migration seemed to occur in some cases, particularly, some clusters in this period migrated to the south as indicated by lined points in the upper panel of Fig. 6b: 1.- April2004-Aug2004 (around day 900), 2.- Sep2005-Aug2006 (1300-1800 days), 3.- Dec2007-Feb2008 (2000-2200 days), 4.- Febr2010-Apr2010 (2900-3000 days), and 5.- Jan2013 (around day 4000). For rainfall, we find insignificant or close-to-zero correlation coefficients on long-term and insignificant results due to high p-values for individual clusters ($0.1<p<0.9$) (Tables 4 and 5). This indicates no correlation with rainfall, neither for long-lasting nor for clustered seismicity. For tides, C-Pearson is almost zero (-0.03) for p-values<0.05 (0.02), which reflects that also tides are not the main driving force for the seismicity in this period (Table 3).

Despite the observed migration of some clusters (Fig. 6b), the results have enough quality to exclude rainfall and tidal variations as significant triggers of the Triacastela seismicity. Having ruled out these processes as potential driving mechanisms, thermal groundwater activity must be considered as the region's relevant trigger mechanism. After 2002, the outcomes reflect a dominance of swarm activity, high b-value (1.2), a low α-value (0.9), and a seismic migration to surface (Table 7, and Fig. 8). Those characteristics have been linked to the effects of fluid migration (e.g., Mogi 1962; Scholz 1968; Hainzl and Fischer 2002; Hainzl et al. 2013; Crespo-Martin et al. 2021). This hypothesis is also supported by: A.-The well-known existence of thermal springs and even crustal fluids in the NW Iberian Peninsula (e.g., Almeida 1982, Pérez et al. 1996; Marques et al. 2003; López et al. 2015), linked to deeply fractured crust and active faults. These studies propose that the excess of mantle helium-3 recorded in the groundwater occurs along the seismogenic faults, representing channels for the potential escape of mantle. The research of Pérez et al. (2008) even points to a high fluid pressure of $CO_2$-rich groundwater as the major role in triggering earthquakes in Triacastela, controlling the nucleation and recurrence of earthquake ruptures. This assumption is supported in the significant increase of earthquake geochemical precursors in this $CO_2$-rich groundwater before the 1995 and 1997 seismicity. B.- The interaction of hydrothermal deep fluids along active faults in this region (González-Menéndez et al. 2022), which caused element mobility by advection in the fault damage zone proximity. These studies show that the faults are permeable channels to deep fluids, and that these hydrothermal fluids percolated producing higher alteration and elements anomalies in the fault surrounding. Our results hold up this prior hypothesis and suggest deep fluids migration as the source of fluid-induced seismicity in the Triacastela.

**7. Concluding remarks**

- The dominance of swarm activity, upward migration of seismicity, the high b-value (1.2) and low α-value (0.9) outcomes after 2002 provide more justification to an hydromechanical coupling, which suggest influence of fluids.

- The Coefficients of Pearson Correlations around zero, for both periods, provide enough quality to exclude rainfall and tides as driven mechanisms of Triacastela seismicity.

- Ruled out these phenomena, the mantellic helium-3 registered in $CO_2$-rich thermal groundwaters in the region, the significant increase of earthquake geochemical precursors in this $CO_2$-rich ground waters before the 1995 and 1997 seismicity, and the hydrothermal alteration and elements mobility through the active faults' planes, allow us to propose deep fluids migration as a possible induced mechanism for long term seismicity in Triacastela.

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**Compliance with Ethical Standards**

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## CAPTIONS

**Fig 1** The study area of Triacastela in the NW Iberian Peninsula (Spain) is marked by the square in the main plot, where red points indicate the recorded earthquakes between 0.3 and 5.1. The inset shows the location of NW Iberian Peninsula (rectangle) within whole Spain, where red points refer to M>4.5 earthquakes in the period from 1988 to September 30$^{th}$, 2018. Triangles are the IGN operated stations on September 30$^{th}$, 2018. Rainfall station of Lugo airport (Stn. Lugo Air) and tide gauge of Vigo (Stn. Vigo2) labeled with purple and coral circles. Focal mechanisms are calculated by the United States Geological Survey – USGS at date May 21$^{st}$, 1997, and time (1) 23:49:45, and (2) 23:50:45 (USGS, 2022)

**Fig 2** (a) Mean annual rainfall in the Iberian Peninsula and Balearic Islands (Spain). Black points are the weather stations. the square indicates the analyzed Lugo station (Source: AEMET. Modified from AEMET 2021b). WGS84. (b) Average monthly precipitation (mm) at the Lugo station between 1995 and 2018 based on the AEMET database

**Fig 3** Oscillation of values for Mean Level, Hide Tide, Low Tide, Max. Tide Range and Min. Tide Range for the beginning to the ending of 2002

**Fig 4** Stacked earthquake monthly activity in the interval (a) 1995 – 2002 and (b) 2002-2008. Gray bars refer to all events, including aftershocks, while the blue bars only refer to background events. The results are compared to randomized data (red line) with its associated variability (plus/minus one standard deviation)

**Fig 5** (a) Earthquakes and tides from 29/11/1995 to 31/12/2001. Earthquakes (red points with y-scale on the left) and Mean Level (grey lines with y-scale on the right) in cm over time. X-axis days relative to 29/11/1995. (b) Earthquakes and tides from 12/01/2002 to 31/12/2018. Earthquakes (red points) and Mean Level (grey lines) in cm over days relative to 12/01/2002 when the first event was registered

**Fig 6** Plots of latitude (x), longitude (y), depth (z), and distance (r) as a function of time for the two periods (a) before 2002 and (b) after 2002. In each case, the distance is calculated to the reference event, first event of each period, marked in pink. In (a) the reference time is 29/11/1995 and the diffusion curve is related to D=0.5 $m/s^2$. In (b), the reference time is 12/01/2002 and the curve is related to D=0.01 $m/s^2$. Clusters with migration patterns are marked by dashed boxes

**Fig 7** Daily earthquakes magnitudes (red points) and rainfall (grey lines). The left y-axis refers to the event magnitudes, and right y-axis refers to rain (mm). (a) Before the network upgrade (from 29/11/1995 to 2002). (b) After the network change (from 01/01/2002 to 2018)

**Fig 8** Annual depth means (black points) and its standard deviation (lines). Stars: mainshocks depth: (1) Mw 4.6 event on 29/11/1995, (2) Mw 4.6 event on 24/12/1995, and (3) Mw 5.1 mainshock on 21/05/1997. Database of depths taking from the relocated catalogue

**TABLES AND CAPTIONS**

**TABLE 1**

| | Formula | $R^2$ |
|---|---|---|
| Mean Level (cm) | y = 0.92x + 5.54 | 0.96 |
| Hide Tide (cm) | y = 0.97x – 3.07 | 0.98 |
| Low Tide (cm) | y = 0.96x – 5.69 | 0.99 |
| Max. Tidal Range (cm) | y = 0.97x + 2.03 | 0.99 |
| Min. Tidal Range (cm) | y = 0.97x + 2.33 | 0.99 |

Table 1. Linear regressions between the Vigo2 station (y-axis) and Villagarcia2 station (x-axis).

**TABLE 2**

| | Winter (%) | Spring (%) | Summer (%) | Autumn (%) |
|---|---|---|---|---|
| 1995-2002 | 14.0 | 44.0 | 18.0 | 5.0 |
| 2002-2018 | 27.1 | 27.3 | 24.9 | 20.6 |

Table 2. Percentage of earthquakes per season for periods before and after 2002. Winter: December, January, and February. Spring: March, April, and May. Summer: June, July, and August. Autumn: September, October, and November. Declustered events are used.

**TABLE 3**

| | | Δt =1 | | Δt =10 | | Δt =20 | | Δt =30 | |
|---|---|---|---|---|---|---|---|---|---|
| Parameters | Period | C-Pearson | p-value | C-Pearson | p-value | C-Pearson | p-value | C-Pearson | p-value |
| Max. Tidal Range | 1995-2002 | 0.02 | 0.26 | -0.02 | 0.73 | -0.05 | 0.60 | 0.09 | 0.43 |
| | 2002-2018 | $-1.0\times10^{-5}$ | 1.0 | -0.02 | 0.68 | -0.07 | 0.22 | -0.10 | 0.14 |
| Min.Tidal Range | 1995-2002 | 0.02 | 0.27 | -0.02 | 0.82 | -0.04 | 0.66 | -0.08 | 0.50 |
| | 2002-2018 | $-2.0\times10^{-3}$ | 0.85 | -0.02 | 0.67 | -0.06 | 0.29 | -0.11 | 0.11 |
| High Level | 1995-2002 | 0.04 | 0.05 | 0.04 | 0.58 | 0.05 | 0.60 | 0.13 | 0.26 |
| | 2002-2017 | -0.01 | 0.35 | -0.04 | 0.33 | -0.09 | 0.13 | -0.13 | 0.07 |
| Low Level | 1995-2002 | $-4.0\times10^{-3}$ | 0.86 | 0.05 | 0.46 | 0.08 | 0.39 | -0.05 | 0.64 |
| | 2002-2017 | $-9.0\times10^{-3}$ | 0.46 | $-3.0\times10^{-3}$ | 0.93 | 0.05 | 0.38 | 0.08 | 0.28 |
| Mean Level | 1995-2002 | **0.05** | **0.03** | 0.09 | 0.17 | **0.20** | **0.03** | 0.08 | 0.50 |
| | 2002-2017 | **-0.03** | **0.02** | -0.08 | 0.05 | -0.06 | 0.27 | -0.11 | 0.11 |

Table 3. Pearson and p-value for correlation between tidal parameters and earthquakes rates (declustered catalogue is used). Tidal parameters are the maximum tidal range (Max. Tidal Range), the minimum range (Min. Tidal Range), the highest (High Level), lowest (Low Level), and mean (Mean Level) values. The results are shown for non-overlapping time bins with alternative bin-sizes of 1, 10, 20, and 30 days. Results with significant p-values (p<0.05) are shown in bold.

**TABLE 4**

| | Δt=1 | | Δt=10 | | Δt=20 | | Δt=30 | |
|---|---|---|---|---|---|---|---|---|
| | C-Pearson | p-value | C-Pearson | p-value | C-Pearson | p-value | C-Pearson | p-value |
| 1995-2002 | $5.0\times10^{-3}$ | 0.80 | -0.03 | 0.67 | $-7.0\times10^{-3}$ | 0.93 | -0.08 | 0.48 |
| 2002-2018 | **-0.03** | **0.01** | -0.07 | 0.10 | -0.06 | 0.30 | -0.06 | 0.43 |

Table 4. Results for rainfall-earthquake correlations in both periods (declustered catalogue is used). C-Pearson: Pearson's correlation coefficient. P-value of $p<0.05$ indicate statistical significance. The approach is tested for non-overlapping time bins of Δt = 1, 10, 20, and 30 days. Significant results are marked in bold

**TABLE 5.**

| | Ƭ=1 | | Ƭ =10 | | Ƭ =20 | | Ƭ =40 | | Ƭ =60 | | Ƭ =100 | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | C | p | C | p | C | p | C | p | C | p | C | p |
| 1995-2002 | $5.0\times10^{-3}$ | 0.80 | -0.01 | 0.59 | 0.01 | 0.50 | 0.02 | 0.40 | -0.02 | 0.45 | $-8.0\times10^{-3}$ | 0.72 |
| 2002-2018 | **-0.03** | **0.01** | $-6.0\times10^{-3}$ | 0.65 | $-7.0\times10^{-3}$ | 0.57 | -0.01 | 0.37 | 0.02 | 0.11 | $-2.0\times10^{-3}$ | 0.88 |

Table 5. Results for rainfall-earthquake correlations in both periods accounting for a potential time delay Ƭ. C: Pearson's correlation coefficient. p: p-value, $p<0.05$ indicates statistical significance. We tested the approach for time shifts between 1 to 100 days, but only Ƭ =1, 10, 20, 40, 60, and 100 days are shown here as a representative sample of results. Significant results are marked in bold. Declustered catalogue is used.

**TABLE 6.**

| Name | Period | Selected cluster from | C-Pearson | p-value |
|---|---|---|---|---|
| October 1997 | 19971019 – 19971218 | Rainfall (49.9mm) | -0.08 | 0.56 |
| September 2001 | 20010922 – 20011122 | Rainfall (62.8 mm) | No events | No events |
| April 2004 | 20040401 – 20040801 | Pore-pressure (1000 days) | -0.14 | 0.12 |
| September 2005 | 20050916 - 20060119 | Pore-pressure (between 1300 and 1500) | 0.03 | 0.71 |
| April 2006 | 20060403 – 20060805 | Pore-pressure (between 1500 and 1800) | -0.05 | 0.54 |
| December 2007 | 20071201 – 20080215 | Pore-pressure (around 2200 days) | -0.19 | 0.10 |
| February 2010 | 20100224 – 20100424 | Pore-pressure and rainfall (56.4 mm) | -0.02 | 0.90 |
| January 2013 | 20130118 – 20130201 | Pore-pressure and rainfall (68.1 mm) | No events | No events |
| December 2017 | 20171210 – 20180510 | Rainfall (63.9 mm) | -0.04 | 0.63 |

Table 6. Pearson correlation for clusters of Triacastela based on pore-pressure (Fig. 6) and rainfall (Fig. 7) observations. Daily binning ($\Delta t$=1) and no time shift ($\tau$ =0) were applied due to the short duration of the clusters.

**TABLE 7.**

| | Mainshock-aftershocks sequences | | | Period |
|---|---|---|---|---|
| | Nov 1995 | Dec 1995 | May 1997 | 2002-2018 |
| Nº events and Mc | 68 events, Mc=2.8 | 67 events, Mc=2.7 | 198 events, Mc=3.0 | 1325 events Mc>1.8 |
| Mogi | Single sq. (1A) | Single sq. (2A) | Single sq. (2A) | Swarm (1C, 2C) |
| α – value (ETAS) | 1.57 | 1.03 | 1.23 | 0.94 |
| b – value (GR) | 1.0± 0.3 | 1.1 ± 0.3 | 1.0 ± 0.2 | 1.2 ± 0.04 |
| p' - value (ETAS) | 1.27 | 1.21 | 1.12 | 1.40 |

Table 7. Outcomes of the main sequences before 2002 (November 1995, December 1995, and May 1997) and the period after 2002. The results of the Mogi Classification (Mogi), Gutenberg-Richter Law (b-value), and Epidemic Type Aftershock Sequence – ETAS fit (the aftershock productivity α – value and the Omori p' -value) are taken from Crespo-Martín et al. (2021) and Crespo-Martín and Martín-González (2021).