

The Castanhão Reservoir, NE Brazil: Reservoir-Triggered Seismicity in secondary transtensional tectonic settings within a major strike-slip system.

Helena Ciechowska¹, Beata Orlecka-Sikora¹, Łukasz Rudziński¹, Aderson do Nascimento², José Augusto Fonsêca¹, Alessandro Vuan^{3,4}

¹*Institute of Geophysics, Polish Academy of Sciences, Warsaw, Poland*

²*Department of Geophysics, Federal University of Rio Grande do Norte, Natal, Brazil*

³*National Institute of Oceanography and Applied Geophysics - OGS, Trieste, Italy*

⁴*National Institute of Geophysics and Volcanology - INGV, Rome, Italy*

Abstract

We investigate Reservoir-Triggered Seismicity (RTS) associated with the Castanhão Reservoir in northeastern Brazil, located within an intraplate region. Using continuous waveform data from a local seismic network (2009–2010), we construct a high-resolution earthquake catalogs of 227 events through template matching (PyMPA) and probabilistic location methods, with subsequent double-difference relocation improving spatial resolution for 103 events. The seismicity forms two clusters delineating a steeply dipping ($\sim 80^\circ$) fault zone characterized as a reactivated strike-slip system with secondary transtensional/stepover structures within the Jaguaribe Shear Zone corridor, with moment magnitudes ranging from M_w 0.0 up to M_w 2.7. To assess triggering mechanisms, we model pore-pressure diffusion induced by reservoir impoundments. The estimated hydraulic diffusivity ($\sim 0.080 \text{ m}^2/\text{s}$) and spatiotemporal evolution of seismicity indicate a response governed by fluid pressure diffusion. Pore-pressure increases on the order of tens of kPa are sufficient to activate critically stressed fault segments, consistent with previous RTS observations. The seismic sequence exhibits swarm-like behavior, interpreted as progressive, fluid-driven rupture within a segmented fault network. We further analyze the temporal evolution of seismic strain using Seismic Strain Dynamics (SSD) and seismic regularity (d_c) metrics, revealing patterns

consistent with a multi-stage earthquake preparatory process. The largest event in the main cluster (M_w 2.1) occurs during a late-stage failure phase, following prolonged fault conditioning. Our results highlight the role of reactivated, critically stressed secondary strike-slip fault networks as hydraulically favorable conduits for pore-pressure diffusion and RTS.. The study demonstrates that even moderate pore-pressure perturbations can induce delayed, swarm-type seismicity in intraplate settings, emphasizing the importance of integrating geological structure, hydromechanical processes, and seismic monitoring for improved hazard assessment near large reservoirs.

Keywords: *Induced seismicity, Earthquake dynamics,*

1. Introduction

1.1. General context

In this paper, we investigate an earthquake swarm triggered by reservoir impoundment, part of the broader class of anthropogenically triggered/induced seismicity generated by human-driven changes in subsurface stress and pore-fluid pressure. However, anthropogenic seismicity can also be related, for e.g., to the exploitation of underground resources or the storage of liquids or gas causing surface deformation (Orlecka-Sikora et al., 2020; Lasocki and Orlecka-Sikora, 2020; Lasocki et al., 2022a; Foulger et al., 2018; Moein et al., 2023). Artificial reservoirs are constructed worldwide for various purposes (e.g., hydropower, irrigation, drought prevention, flood control, etc.). Regardless of their purpose, large artificial lakes can trigger seismic activity (Gupta, 2002). Reservoir-Induced Seismicity (RIS) was first documented in the Boulder Dam area, at Lake Mead, USA, by Carder (1945), who linked it with reservoir water levels. RIS, such as in this case, can produce earthquakes of magnitude ~ 4 that can be felt locally, causing unrest and anxiety among residents of affected areas. Strong earthquakes can cause damage, with catastrophic effects on nearby cities' infrastructure, the quality of life, and even fatalities.

The terms Reservoir-Induced Seismicity and Reservoir-Triggered Seismicity (RTS) are often used interchangeably (Gupta, 2011). The problem of terminology was addressed by McGarr and Simpson (1997), who proposed a distinction between “induced” and “triggered” reservoir-related earthquakes. According to McGarr et al. (2002), triggered seismicity occurs when an external perturbation initiates the nucleation of an earthquake in a small region, accelerating its occurrence. In this case, even a small change in stress caused by a triggering phenomenon causes the seismic event which would happen anyway. In contrast, induced seismicity occurs when an external process, such as fluid injection or underground mining, significantly changes the local regional stresses leading to fault failure that would not occur in the absence of such perturbations. According to Gupta (2011), earthquakes occurring after reservoir filling are commonly categorized as “triggered” since the pore pressure changes related to reservoir filling do not significantly alter the residual stress field (do Nascimento et al., 2005; Orlecka-Sikora et al., 2023; Nowaczyńska, 2025). For this study we use the term RTS. A better understanding of this phenomenon is crucial for public safety and for maintaining confidence in critical infrastructure related to hydropower and flood control.

History shows that RTS can pose a significant risk to human life (e.g.: the 1967 Koyna earthquake in India, which remains the strongest recorded RTS event to date). The damage from this M_w 6.3 earthquake included destruction resulting in ~200 fatalities and ~1,500 injuries. Infrastructure was destroyed, leaving thousands of people homeless (Gupta, 2002). Such events underscore the importance of studying RTS to prevent tragic consequences by better understanding this phenomenon.

Seismicity within the artificial reservoirs can occur as a mainshock-aftershock sequence (Hainzl et al., 2016) or as a swarm (Hainzl, 2003; Hainzl, 2004; Orlecka-Sikora et al., 2023; Valoroso et al., 2023; Talwani, 1997; Grasso et al., 2018). A swarm is a seismic sequence confined to a limited space and time interval without a clearly dominant mainshock, with events of comparable magnitude. Recent studies suggest that seismic swarms can reflect intermittent rupture processes, in which seismic moment is released progressively through multiple small events, while deformation and damage accumulate on the fault system rather than being

accommodated by a single large rupture (e.g., Orlecka-Sikora et al., 2023, Palgunadi et al., 2024). In this framework, swarm activity indicates distributed fault weakening and deformation that may either dissipate energy without system-scale failure or, under specific conditions, contribute to the organization of deformation toward larger earthquakes.

The Castanhão Reservoir is located in northeastern Brazil (Figure 1), in one of the most seismically active regions within the stable continental part of South America (Ferreira et al., 1998; Bezerra et al., 2011). While Brazil has a documented history of RTS, with approximately 30 cases identified nationwide and a general understanding of the primary triggering mechanisms (loading and pore pressure changes) (Assumpção et al., 2002; do Nascimento et al., 2004, 2005; Barros et al., 2018; Raza et al., 2023; Sayão et al., 2020), significant knowledge gaps remain, particularly in relations to detailed, integrated case studies.

Comprehensive investigations that rigorously link observed seismicity patterns to specific reservoir characteristics, operational histories, detailed subsurface geology, hydrogeological properties, and the local stress field are relatively scarce across Brazil (Silva et al., 2020; do Nascimento et al., 2004, 2005, Barros et al., 2018). Dedicated, long-term monitoring campaigns explicitly designed for RTS analysis have been limited, particularly given the large number of reservoirs constructed over the decades, many of which predated modern monitoring capabilities (Bianchi et al., 2018; Do Nascimento et al., 2004; Ferreira et al., 1995). Even in NE Brazil, despite its known natural seismicity and relatively better monitoring coverage provided by the Brazilian Seismographic Network (RSBR) (Bianchi et al., 2018), correlating seismicity (whether natural or potentially induced/triggered) with specific geological structures remains challenging (Assumpção et al., 2014; Bezerra et al., 2011). Therefore, a focused study of the Castanhão Reservoir presents a valuable opportunity.

The data available for the Castanhão enables a more detailed analysis than was previously possible for many older reservoirs that suffered from historical data gaps (Bianchi et al., 2018). By integrating seismic observations with reservoir parameters and the specific geological setting of NE Brazil, this study directly addresses the identified need for more well-characterised and theoretically understood RTS examples in Brazil (Sayão et al., 2020). It

offers the opportunity to contribute crucial data points to the limited national dataset, potentially leveraging regional seismotectonic knowledge and enhancing monitoring infrastructure.

Here, we describe the construction of the earthquake catalogue, from initial detection and probabilistic location through double-difference relocation and moment magnitude estimation, followed by the formulation of the pore-pressure diffusion model applied to the reservoir's impoundment history. Then, we present the spatial and temporal characteristics of the relocated seismicity, the results of the pore-pressure modelling, their relation to the regional structural setting, and the temporal evolution of seismicity assessed through the Seismic Strain Dynamics and seismic regularity metrics. At the end, we summarize the main conclusions and their implications for reservoir-triggered seismicity hazard assessment, together with the limitations of the present study.

1.2 The Castanhão region description

The study area is located in the state of Ceará, NE Brazil. The Castanhão Reservoir was created in 2002 following the construction of the dam on the 610 km-long Jaguaribe River. The Castanhão Reservoir is the largest water reservoir in the State. Its storage capacity is 6.7 billion cubic meters (Rocha et al, 2011). According to Camarão Jr. (2001), the reservoir has a surface area of up to 325 km². The construction lasted 13 years (1990-2003) and resulted in a 60 m-high dam (Nunes et al., 2011).

1.2.1. Geology

The geology of the Castanhão Dam inundation area can be described by at least five lithostratigraphic units. In 2001, Camarão Jr. identified them as a gneissic-magmatic basement, a metavolcanic-sedimentary sequence, Brazilian granitoids, swarms of Mesozoic basaltic dikes, and Cenozoic fluvial terraces of the Jaguaribe River, over which the dam was constructed. Subsequent investigations by Ferreira et al. (2008) described the area mainly as Precambrian crystalline bedrock and consisting mostly of metamorphic rocks, including gneisses, migmatites, micaschists, and paragneisses. These units are intruded by Precambrian granitic rocks. A basalt dike swarm cuts the crystalline basement, and the

crystalline rocks are capped by 5 – 10 m thick Late Quaternary deposits younger than 0.5 Ma (Ferreira et al., 2008).

1.2.2. Tectonics

According to Ferreira et al (2008), there are two principal groups of faults: (1) faults formed at ductile-brittle transition crustal level and (2) shallow faults (depth less than 4 km). Group (1) is associated with the cataclasites and epidotes, while the fault breccia and gouge form the (2). Group (2) cuts through all identified geological units, suggesting that these structures may be of late Quaternary age. Both groups are described as high-angle strike-slip structures, and most exhibit predominantly low-rake ($\leq 15^\circ$) strike. However, a dip-slip is evident in certain places. The faults exhibit either a left-lateral or right-lateral sense of shear. N–S and NE–SW trending faults are related to ductile Precambrian fabric reactivation (Ferreira et al., 2008). Figure 1 shows the map based on local tectonics mapped by Cavalcante et al. (2003).

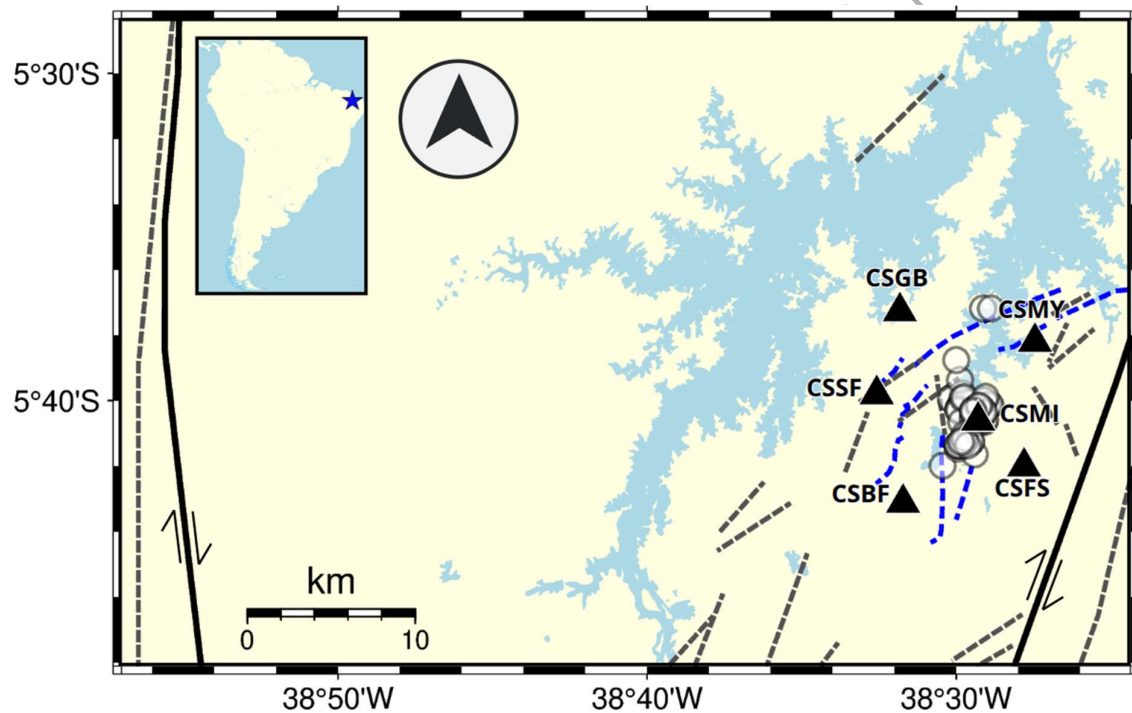


Figure 1. Map of local faults in the Castanhão Reservoir region, adapted from Cavalcante et al. (2003). Solid black lines indicate the principal mapped faults, while dark-gray dashed lines mark secondary fractures. Blue dashed lines delineate additional lineaments interpreted from the Digital Elevation Model (DEM). Light blue shaded areas correspond to surface water coverage,

while light yellow areas correspond with land. Black triangles denote the stations of the LS seismological network. White circles with gray outlines show the initial earthquake locations obtained using the TRMLOC algorithm, which served as templates for the subsequent matching procedure applied in this study.

1.2.3. Seismicity

Direct observations of local background seismicity in the vicinity of the reservoir prior to impoundment are not available. However, previous studies (Assumpção et al., 1989; Ferreira et al., 1998; Fonsêca et al., 2021) document the seismic activity within Cascavel-Pacajus (~50 km north of Castanhão Reservoir), Palhano (~100 km north of Castanhão Reservoir), and Tabuleiro Grande (~35 km southeast of Castanhão Reservoir), which are situated around the Potiguar basin and are close to the study site.

Seismic activity in Cascavel county has been documented since 1980, when the largest, 5.2 m_b Pacajus earthquake struck on November 20, causing damage to the local infrastructure. Its aftershocks were recorded for over two years (Ferreira et al., 1998). Following the Pacajus earthquake, 749 seismic events occurred between 1989–1991. Later, between October and November 1993, earthquakes of 2.0 m_R occurred, causing damage to a nearby water channel. Seismicity intensified in June 1994, culminating in the 3.3 m_R event on 11 August 1994. Low-magnitude earthquakes persisted until April 1998 (Ferreira et al., 1998). Between November 2016 and December 2017, the seismic activity increased once more. Small events (~3.0 m_R) were recorded until the end of 2017 (Fonsêca et al., 2021).

In December 1987 the Palhano swarm occurred, with its increase in October 1988 and an earthquake of 4.2 m_b on October 19th, 1988 (Assumpção et al., 1989; Ferreira et al., 1998). From that day, until almost the end of June 1993, 28393 seismic events were recorded, with the largest event (4.5 m_b) on March 27th, 1989 (Ferreira et al., 1998). Moreover, 13 shallow (depths between 1.0 and 2.5 km) earthquakes were recorded 6 km west of Tabuleiro Grande (Ferreira et al., 1998).

Studies on seismicity in the Castanhão Reservoir area date back to 2003, but at that time the measurements were discontinuous. According to Ferreira et al. (2008), it is not possible to

determine when the RTS activity began, as the first earthquake within the vicinity of the Castanhão Reservoir was recorded in 2003, when the seismological network was deployed. Between March and April 2003, there was at least one operating station. In April 2003, the network of four seismological stations became available (one analogue and 3 digital stations). The recorded seismic events are not available for inspection. The same studies showed that seismicity occurred in clusters, each active during different periods - the authors focused on seismicity between March 2003 and April 2005 and estimated that event magnitudes were mostly below 1.0 m_b , with a maximum of 1.4 m_b (Ferreira et al., 2008). The same research followed with a more detailed analysis of 2004 seismological data, which resulted in the conclusion of RIS due to reservoir filling- and heavy rainfall season-related pore pressure change. No seismological data were collected between April 2005 and November 2009. Later, Nunes et al (2011) analyzed the Castanhão Reservoir seismicity from November 2009 to December 2010, identifying over 500 seismic events grouped within three clusters that activated in different periods. The hypocentral positions of these earthquakes aligned with the faults mapped by Camarão Jr. (2001). However, the magnitudes of the identified events were not reported. Unfortunately, none of the catalogs described by other researchers (Ferreira et al., 2008; Nunes et al., 2011) is available. Consequently, in this study we constructed a new seismic catalog for the Castanhão Reservoir by scrutinizing available continuous waveforms from six seismological stations of the local network (LS) operating between November 2009 and December 2010. We further estimate the moment magnitudes and investigate the possible triggering mechanism of the observed swarm, as well as evaluate the effects of pore-pressure changes associated with reservoir impoundment on local seismicity.

2. Methodology

2.1. Preparation of the earthquake catalog

The data were recorded at 6 seismological stations deployed southwest of the reservoir. The network distribution is shown in Figure 1. The dataset consists of three component recordings of continuous waveforms from mid-November 2009 to the end of December 2010 – the same dataset was analyzed by Nunes et al. (2011). The instruments used for seismic monitoring were six triaxial short-period Sercel L-4-3D 2Hz seismometers, WO/CT, 5500 Ohms, distributed in a radius of about 10 km. During network operation, some stations were intermittently offline, and their settings were occasionally modified (e.g., sampling rate – see Supplementary Materials). To maintain dataset consistency within the selected time frame, we used only recording channels with a sampling rate of 250 samples per second. In addition, the limited station coverage makes it difficult to determine earthquake focal mechanisms. Only six stations are available, one of which (CSMI) presents inconsistent signal shifts making its data unreliable.

For the initial search for seismic events, the STA/LTA method was applied for 2010 data, resulting in 53 templates with suitable signal-to-noise ratios (S/N). The template set was later expanded to 69 by incorporating additional events from the 2009 data. Each selected earthquake was located using the TRMLOC algorithm on the European Plate Induced Seismicity Observations and Datasets within the EPOS Services (EPISODES) Platform (Dębski, 2015; Orlecka-Sikora et al, 2020; Lasocki et al, 2022b). TRMLOC locates hypocenters by inverting first arrival times using the probabilistic inverse theory. Maximum likelihood hypocentral solutions were obtained and utilized for further analysis. These results formed the initial catalog of template events used for cross-correlation detection with the Python Matching Phase Algorithm (PyMPA; Vuan et al., 2018). PyMPA is an open-source Python package designed for detecting seismic events. The algorithm is based on cross-correlating continuous waveform data with templates. It utilizes the ObsPy package (Krischer

et al., 2015) to handle both data and their metadata. The earthquakes included in the initial PyMPA catalog are shown in Figure 1.

For the following analysis, PyMPA was configured to use a frequency range of 5-100 Hz. Because locating a seismic event requires records from at least three stations, only events meeting this criterion were retained for further analysis. Since the CSMI station experienced frequent issues such as shifts in the signal after some service interruptions, continuous waveforms were available for only 5 stations (15 channels).

PyMPA detections required an average cross-correlation of 0.35 across at least six channels. This choice maximized the number of detected events ensuring a proper S/N ratio (avoiding triggering false detections). At this stage, PyMPA detected ~450 events. After visual inspection (clear P-wave phase at least on 3 different stations) 227 met the quality criteria for further analysis.

These 227 events were located with the Hypo71 software (Lee and Lahr, 1972). As expected, depth estimates exhibit significant errors (mean vertical error of 6.2 km); therefore, we relocated the Hypo71 catalog using the HypoDD algorithm (Waldhauser and Ellsworth, 2000; Waldhauser, 2001). The double-difference method allows relative relocation based on three types of input: existing catalog locations (catalog-based relocation), cross-correlation between event waveforms (cross-correlation-based relocation), or a combination of the two. For this study, we applied the combined Double Difference method, which allowed us to relocate 103 earthquakes. The remaining 124 events could not be relocated because they were too distant from neighboring events. Because HypoDD produced unrealistically small formal errors (meters to tens of meters), we recalculated the error using the bootstrap method with 95% confidence interval, reducing the mean vertical error to ~850 meters (Supplementary Materials).

For both Hypo71 and HypoDD catalogs we used the trimmed seismic signals to estimate moment magnitude M_w and to attempt computation of focal mechanisms. This analysis was done using Pyrocko (Heimann et al., 2017) and KiwiTools software (Cesca et al., 2010).. The full waveform inversion with KiwiTools was obtained by using a local velocity model, with the assumption of a homogenous medium, with the P-wave velocity of $V_p = 5.95$ km/s (Nunes et

al., 2011), and $V_p/V_s = 1.74$ (Supplementary Materials). We conducted analysis in the frequency domain, yielded moment magnitude (M_w) estimates for each event. The completeness magnitude, $M_c = 0.4$, of each catalog was estimated using the tools available on the EPISODES Platform (Orlecka-Sikora et al, 2020). The M_c was estimated by the Goodness of fit test at 90% confidence bounds (Wiemer and Wyss, 2000; Leptokaropoulos et al., 2013). Due to the quality of the data, it was not possible to compute M_w for all of the earthquakes, hence some of the events lack the defined magnitude (7 events do not have their M_w values assigned). Due to the data quality and the limited number of stations, it was not possible to obtain stable focal mechanisms for seismic events.

To obtain the most complete earthquake catalog, we merged Hypo71 and HypoDD outputs into one catalog, replacing successfully relocated Hypo71 events from the initial catalog with their HypoDD-relocated equivalents. For further analysis, we use this combined catalog. Despite the larger hypocentral location errors associated with the Hypo71 solutions (discussed in detail in Supplementary Materials), their inclusion in the combined catalog provides sufficient location accuracy for the following analysis while allowing us to retain the largest possible number of events. After merging the catalogs into a single combined one, 8 events lack M_w estimates. Additionally, we estimated the b -value for local seismicity as $b=1.29$, while the regional b -value reported by Fonsêca et al. (2024) ranges from 0.82 to 0.94. According to Mallika et al. (2013) b -values associated with RTS are higher than regional ones, which seems consistent with our observations.

2.2. Pore pressure computation

To further investigate the Castanhão Reservoir seismicity, we calculate the pore pressure change within it using Orlecka-Sikora et al. (2023) methodology. We computed the hydraulic diffusivity using formula (1) as introduced by Talwani et al. (2007):

$$c = \frac{L^2}{4t}, \quad (1)$$

where the diffusion c is estimated from time delay t and the distance L between the earthquake source and the pressure front origin. In this case we defined L as a distance between

earthquake hypocenters and reservoir borders. This approach was adapted because the reservoir is a spatially scattered source and the actual hydraulically connected flow pathways are unknown. Such assumption provides a reproducible lower-bound estimate of possible propagation distance. The derived diffusivity should be treated as effective diffusivity under a minimum-path assumption, rather than a direct estimate of the intrinsic hydraulic diffusivity of the medium. This assumption provides a geometric approximation corresponding to the shortest possible hydraulic pathway between the reservoir and each earthquake hypocenter. For comparison, we also estimated using the distances between earthquakes and reservoir center, as well as the interevent distances; these alternative approaches are further discussed in the Supplementary Materials). The closest-point approximation applies only to the estimation of hydraulic diffusivity, whereas the subsequent pore-pressure modelling accounts for the contribution of the whole reservoir. After computing diffusivity c for all events in the catalog, we calculated mean diffusivity and its confidence intervals that were later used for modeling of pore pressure changes.

Following Kalpna and Chander (2000), the evolution of pore pressure spatiotemporal changes was solved throughout a half-space using the Green's function approach with the determined initial conditions and boundary. For our study we assumed the grid of $\Delta x = \Delta y = 50$ m, and $\Delta z = 40$ m. The time step in our calculations was assumed as 30 days, starting from the reservoir filling on January 2nd, 2004. We used the homogenous diffusion formula (2) for porous elastic rocks (Rice and Cleary, 1976):

$$c\nabla^2\Delta p - \frac{\partial}{\partial t}\left(\Delta p - \frac{B}{3}\Delta\theta\right) = 0, \quad (2)$$

where $\Delta\theta$ defines the change in mean stress, B defines the Skempton's coefficient, and Δp is pore pressure change. Following empirical studies by Mesri et al. (1976) and Zhang et al. (2009), Skempton's coefficient was assumed to be 1.0. The solution to the above formula (2) for the diffusion component of the pore pressure variability was defined by formula (3):

$$\Delta p_d(x, y, z, t) = c \int_{-\infty}^{+\infty} \int_{-\infty}^{+\infty} r(x', y') \cdot f(t') \frac{\partial G}{\partial z'} \Big|_{z'=0} dx' dy' dt', \quad (3)$$

where x, y, z denotes the location at which the pore-pressure change is evaluated and $(x', y', 0)$ denotes a point within the loaded surface area of the reservoir. G represents a Green's function (Kalpna and Chander, 2000). This methodology was discussed in detail in Supplementary Materials for Orlecka-Sikora et al. (2023).

3. Results and Discussion

3.1. Spatial patterns

The following analysis was performed on a final catalog of 219 events. Figure 2 shows this combined catalog in the 3D volume (three plots on two-dimensional projections).

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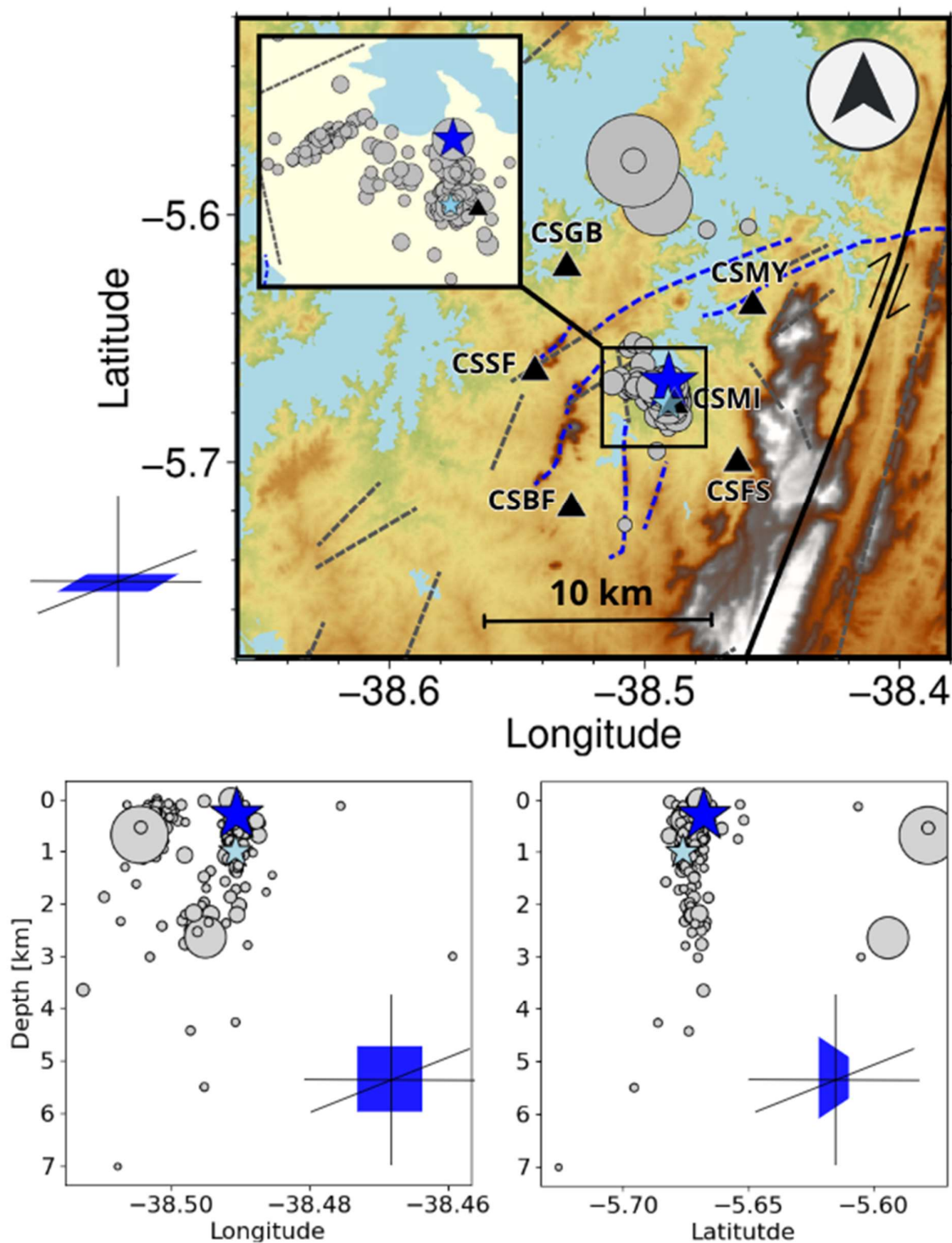


Figure 2. Spatial distribution of seismic events from the combined catalog (gray circles, scaled by moment magnitude M_w 0.0–2.7). The light blue star represents the first earthquake in the swarm, while the blue star denotes the strongest M_w 2.1 event in the cluster. Black triangles indicate the LS network seismological stations. The solid black line marks the major mapped

faults, while dark-gray dashed lines represent fractures identified by Cavalcante et al. (2003).

Blue dashed lines show additional tectonic lineaments interpreted from the Digital Elevation Model (DEM). The panels illustrate the three-dimensional geometry of the seismicity projected onto: **(top)** longitude–latitude over the topographic map with a close up on two clusters, **(bottom left)** longitude–depth, and **(bottom right)** latitude–depth.

Two main seismic clusters can be identified based on their spatial distribution. In addition, several events occur outside them, potentially indicating a third, less-defined cluster located north of the main seismicity area. This northern cluster is not resolved in the present study but was reported by Nunes et al. (2011). Two relatively strong earthquakes (M_w 2.3 and M_w 2.7) were located north of the LS network.

The relocated spatial distribution of two clusters apparent in map view and depth (Figure 2), delineates a distinct, steeply dipping fault geometry (approximately 80° dip) and NE–SW and N–S strikes. In the context of two major subparallel strike-slip faults of similar kinematics in the study area that can be seen in Figure 1, the intervening steeply dipping deformation zone imaged by seismicity is consistent with a secondary transtensional structure.

Focal mechanism solutions could be computed for only a few individual events within the clusters using KiwiTools and were poorly constrained due to limited waveform fits. To further explore the faulting style, composite focal mechanisms were therefore calculated; however, the available data impose limitations on the robustness of these results and they should be treated as indicative rather than definite solutions (Supplementary Materials).

The interpreted fault zone extends approximately 2 km along strike and 3.5 km downdip, for an estimated rupture area of about 7 km^2 . The length was measured as the epicentral distance between the most northern and southern event in the N–S striking cluster, while the downdip was assumed based on the depths where earthquakes occur. Using the scaling relations for strike-slip faults from Wells and Coppersmith (1994), the rupture area (A , in km^2) and moment magnitude (M_w) are related by:

$$\log_{10}(A) = -3.42 + 0.90 \cdot Mw (\sigma = 0.22). \quad (4)$$

Applying this relationship to the rupture area yields a potential earthquake magnitude of $Mw \approx 4.7$. Similarly, regressions for subsurface rupture length (RLD) and downdip rupture width (RW) give:

$$\log_{10}(RLD) = -2.57 + 0.62 \cdot Mw (\sigma = 0.15), \quad (5)$$

$$\log_{10}(RW) = -0.76 + 0.27 \cdot Mw (\sigma = 0.14). \quad (6)$$

These estimates suggest potential magnitudes of $Mw \approx 4.6$ (length) and $Mw \approx 4.8$ (width). The variability reflects both the empirical uncertainties of the regressions and the significant errors inherent in depth estimation, which strongly affect width-based scaling. The observed seismicity is limited to events of Mw 2.3–2.7, which are significantly smaller than the maximum earthquake size predicted from the inferred fault dimensions. This indicates that the identified fault structure may be capable of hosting larger earthquakes than those observed during the period covered by the available catalogs.

The two main seismic clusters may indicate the activation of multiple fault strands or echelon segments – discrete, overlapping fractures oriented obliquely to the regional fault trend, and commonly associated with strike-slip systems affected by a component of extension.. Deeper, scattered events may reflect fluid-assisted failure or off-fault damage. The observed activity can be interpreted as a spatially and temporally distributed rupture process involving the progressive activation of the fault segment. Unlike a typical tectonic earthquake, in which rupture propagates dynamically over seconds, the inferred fluid-driven rupture developed progressively through a sequence of individual earthquakes forming a seismic swarm over approximately one month.

It is not clear whether any events occurred before December 2009, as continuous data for the earlier period are unavailable (Figure 3). The full data availability for each station of the LS network was shown in Supplementary Materials for this work.

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3.2. Pore pressure modelling

Based on previous studies (Ferreira et al., 2008), the recorded seismicity is consistent with RTS. Such activity was observed and found to be correlated with reservoir water level by researchers investigating the Castanhão Reservoir earlier (Ferreira et al., 2008; Nunes et al., 2011). In the case of Ferreira et al (2008) studies the main argument for an RTS was the abrupt increase in seismicity in early 2004 soon after the rapid increase in water level. We applied the pore pressure diffusion model (Orlecka-Sikora et al., 2023) to the combined catalog for a better understanding of the triggering mechanism. Because seismological observations are not available from the initial stage of reservoir impoundment, the timing of reservoir loading could not be used as a reference. However, we still tested the assumption for a case in which the first earthquake occurred at the beginning of reservoir filling in 2002 at the location of cluster “A” identified by Ferreira et al. (2008) in which computed mean diffusivity was $0.004 \text{ m}^2/\text{s}$ (95% CI: $0.002 \text{ m}^2/\text{s} - 0.006 \text{ m}^2/\text{s}$). These results are discussed in detail in Supplementary Materials. For this work the first recorded earthquake in the catalog was assumed to occur at time $t = 0$. Time delays for subsequent events were measured relative to this reference event, whereas distances were calculated as the shortest hypocentral distance to the reservoir. Seismic hydraulic diffusivity was then estimated for each earthquake assuming diffusion-controlled triggering, following the distance–time scaling proposed by Talwani et al. (2007). In this framework, the resulting diffusivity values represent effective (apparent) diffusivities characterizing the observed seismic sequence. Some diffusivity estimates yield unrealistically high, due to short inter-event times. To obtain diffusivity values representative of the local crystalline basement, the analysis was therefore restricted to diffusivities $\leq 10 \text{ m}^2/\text{s}$, retaining 207 out of 227 events. This upper limit is consistent with values reported for fractured crystalline rocks (Talwani et al., 2007; Orlecka-Sikora et al., 2023). The resulting mean hydraulic diffusivity is $0.080 \text{ m}^2/\text{s}$ (95% CI: $0.061 - 0.100 \text{ m}^2/\text{s}$). For comparison, do Nascimento et al. (2005) reported diffusivities of $\sim 10^{-3}$ to $10^{-2} \text{ m}^2/\text{s}$ in the Açú reservoir in NE Brazil, while Parotidis et al. (2003) and Scholz (2002) documented diffusivities up to $10 \text{ m}^2/\text{s}$ in fault zones. More recently, Orlecka-Sikora et al. (2023) estimated a mean diffusivity

of $0.24 \text{ m}^2/\text{s}$ (95% CI: $0.06\text{--}1.0 \text{ m}^2/\text{s}$) for triggered seismicity at the Song Tranh 2 Reservoir in Vietnam. The inferred diffusivity values suggest that pore-pressure diffusion governed the delayed seismic response and that seismicity initiated once the diffusing pore-pressure perturbation reached a pre-existing, critically stressed fault segment near the reservoir.

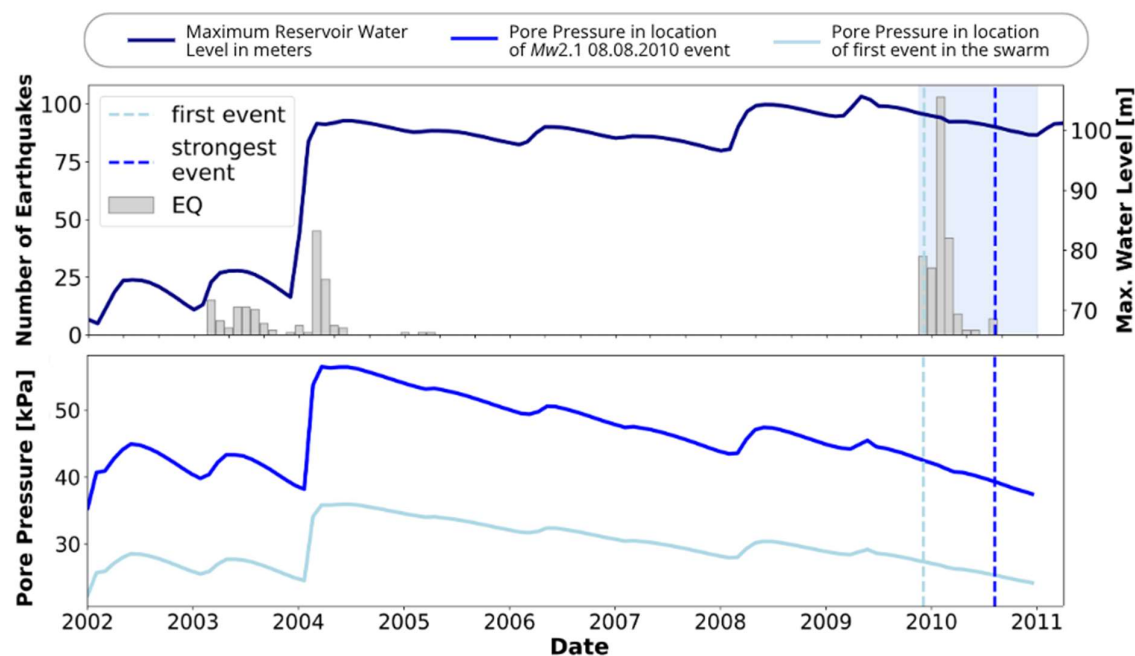


Figure 3. Seismic activity alongside maximal reservoir water levels (given in meters), and pore pressure variability at the hypocentral location of the largest event in the main cluster. The dashed blue line marks the time of $M_w2.1$ main event, while the light blue dashed line marks the moment of first event within the investigated swarm; **(Top)** Monthly seismic activity compared with the maximum water level of the Castanhão Reservoir. The seismicity in 2003 – 2005 was adapted from the plots in Ferreira et al. (2008). The blue-shaded region indicates the time during which LS network data were available. The solid navy-blue line shows the monthly maximum reservoir water level (in meters). Light-gray bars represent the number of seismic events per month from the Hypo71-located earthquake catalog; **(Bottom)** Pore-pressure evolution between 2002 and 2011. The light-blue curve denotes pore-pressure changes at the hypocenter of the first earthquake in the Castanhão swarm (4

December 2009). The dark-blue curve shows pore-pressure changes at the hypocenter of the M_w 2.1 event—the largest earthquake in the cluster, which occurred on 8 August 2010.

The available seismic observations do not allow the onset time of the seismic response to reservoir impoundment to be reliably determined. In particular, the apparent onset of seismicity may reflect the temporal coverage of the monitoring rather than the actual onset of the reservoir-related seismic response. Nevertheless, the available reservoir water-level history and the locations and occurrence times of the recorded earthquakes allow us to examine the hydromechanical conditions associated with the observed seismicity. For this purpose, we calculated the evolution of the reservoir-induced pore-pressure perturbation at the earthquake locations. The pore-pressure evolution was calculated for the spatially distributed load of the entire reservoir employing the method proposed by Orlecka-Sikora et al. (2023). The solution is based on the diffusion equation and the Green's function formulation of Kalpna and Chander (2000), accounting for the reservoir geometry and its time-varying water load. The measurement period was divided into finite time steps, and the water load was discretized into a set of point forces distributed over the reservoir surface. The solution accounts for both instantaneous and time-dependent pore pressure changes due to varying hydrostatic loads. Pore pressure values were evaluated at the hypocentral coordinates of the seismic events, and at their respective times of occurrence (Figure 3 and Figure 4). The model assumes a homogenic medium, whereas the actual rock mass is fractured and structurally heterogeneous. Preferential hydraulic pathways and variations in hydraulic properties may therefore produce local pore-pressure conditions that differ from the modeled values.

The modeled pore-pressure evolution does not simply follow the absolute reservoir water level. The rapid increase in reservoir level in 2004 produces the largest increase in modeled pore pressure, whereas during the following years the pressure generally decreases despite the persistently high reservoir level, with shorter-term variations superimposed on this long-term trend. This behavior reflects the time-dependent response to the history of reservoir loading rather than a direct relationship with the instantaneous water level. The seismicity

recorded in 2009–2010 occurred during this declining phase of the modeled pore-pressure evolution.

The mean total pore-pressure change at the location of the main seismicity cluster and at the time of the *Mw*2.1 earthquake is approximately 40 kPa, combining contributions from boundary-induced transient effects, poroelastic diffusion, and the static Boussinesq response. In contrast, the modeled pore-pressure perturbation at the location of the largest earthquake in the catalog (*Mw*2.7, 14 April 2010) is substantially smaller throughout the analyzed period, ranging from 0.055 to 0.095 kPa, with a mean value of 0.086 kPa. The substantial difference between the modeled pore-pressure perturbations at the two locations cannot be explained by their minimum distance from the reservoir boundary, as the two earthquakes are located at comparable minimum distances from the reservoir (Figure 4). The pore-pressure model accounts for the spatially distributed load of the entire reservoir; therefore, the calculated pressure at a given hypocenter depends on its three-dimensional position relative to the whole loaded reservoir area and on the temporal evolution of the reservoir water load. The different modeled pore-pressure histories at the two hypocenters thus reflect the combined effects of their spatial relationships to the distributed reservoir load and the evolution of that load through time.

The occurrence of the *Mw*2.7 earthquake under a much smaller modeled pore-pressure perturbation does not necessarily exclude a hydraulic contribution to triggering.

In critically stressed intraplate crust, transient pore-pressure increases as low as 0.1–0.5 kPa are sufficient to trigger shear failure on favorably oriented faults, as previously demonstrated at the Açú Reservoir (do Nascimento et al., 2005). The contrast between the two Castanhão locations suggests that the relationship between modeled pore-pressure perturbation and earthquake magnitude is influenced by local stress and fault conditions.

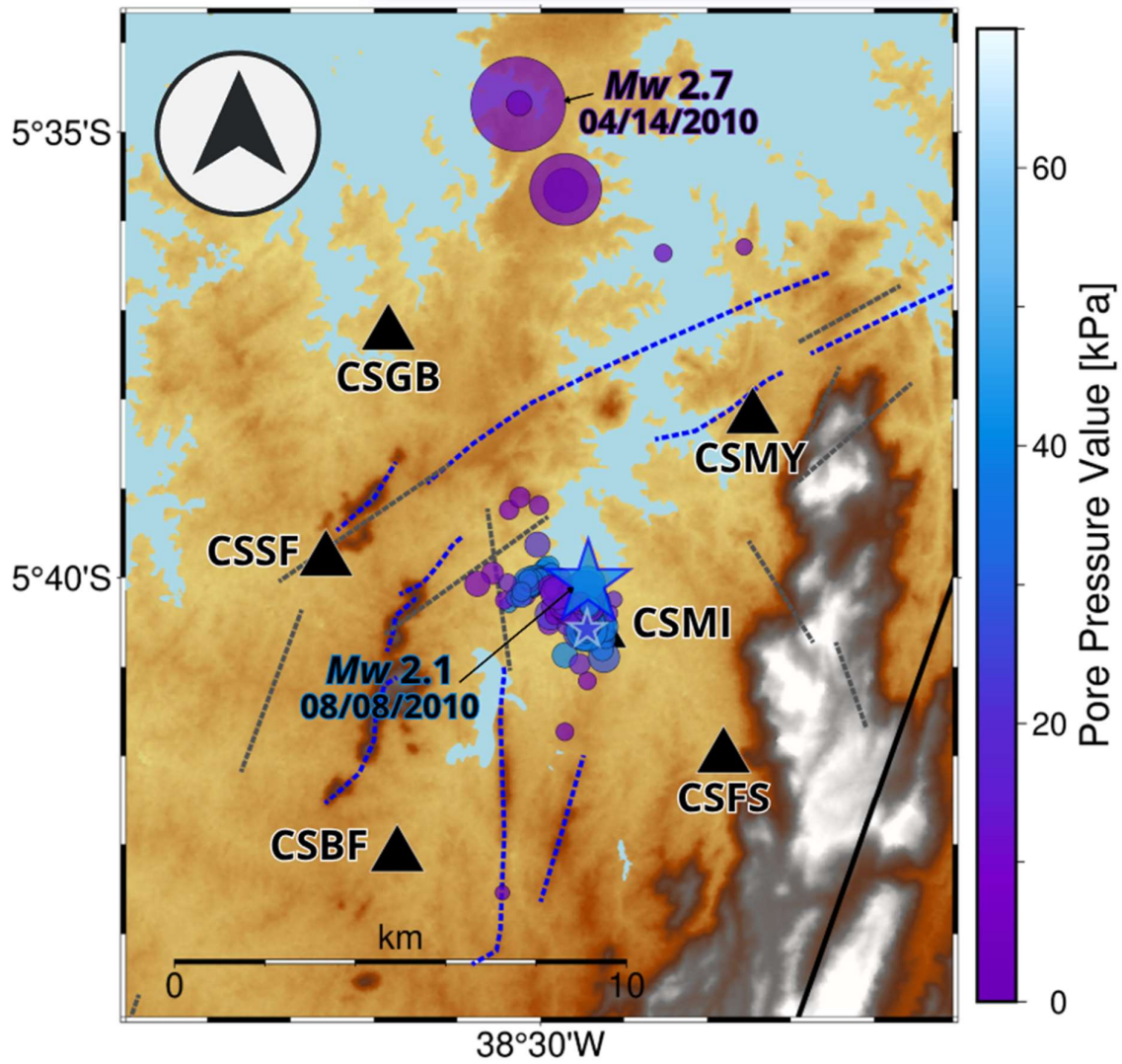


Figure 4. Pore pressure at earthquake hypocenters on the day of the *Mw* 2.1 event (8 August 2010). Calculations are based on the combined seismic catalog and are displayed over a topographic map. Circles colors indicate the total pore-pressure values (in kPa), while point sizes scale with event magnitude, stars mark the first (light blue outline) and main event (blue outline) in the cluster. Black triangles denote the locations of the LS network seismological stations.

3.3. Secondary transtensional structures as critical hosts for Reservoir-Triggered Seismicity.

The tectonic setting of the area around the artificial lakes can strongly influence the spatial distribution and character of RTS. A consistent pattern emerges across several RTS case studies – including the Açú Reservoir (do Nascimento et al., 2005), Song Tranh 2 (Orlecka-Sikora et al., 2023), the Reykjanes Peninsula (Hrubcová et al., 2021), and our study of the present Castanhão Reservoir. These highlight the central role of secondary transtensional fault systems in governing triggered seismic response. Such structures (e.g. pull-apart systems) are typically formed in transtensional regimes and may comprise segmented, en echelon fault arrays. They are characterized by steep dips (70–90°), mechanical weakness, and enhanced hydraulic heterogeneity. Pre-existing fractures and fault intersections may provide preferential pathways for fluid migration and localize reservoir-induced stress and pore-pressure perturbations, making these structures particularly susceptible to reactivation when favorably oriented within the local stress field. Seismicity in such a system is commonly described as swarm-like. There is no clear mainshock-aftershock sequence, and the earthquakes are often spatially organized in two opposing fault segments or clusters, inclined toward each other. Such geometry promotes progressive failure via a diffusive rupture front. The spatial organization of the Castanhão seismicity is consistent with activation of such a structurally complex fault system. The two principal seismic clusters may represent activity on interacting fault strands or en echelon segments, while the deeper and more scattered events may reflect activation of subsidiary structures or off-fault deformation. The observed sequence has a swarm-like character, with seismic activity distributed between different parts of the fault system. Within the available 2009–2010 catalog, the swarm sequence developed gradually and culminated in the relatively strong event ($M_w 2.1$) at the end of the cluster activity. This spatial and temporal organization is consistent with progressive activation of different parts of a heterogeneous fault network rather than rupture of a single fault plane in one dynamic event. Similar behavior has been observed in other structurally complex fault systems. In the Narlı region, slow deformation and seismic bursts preceded a larger earthquake, illustrating how

deformation may be distributed among interacting secondary structures before larger-scale rupture (Karabulut et al., 2025).

A further example is provided by RTS in the Koyna–Warna region, India. The largest RTS earthquake ($M_w \approx 6.3$) occurred several years after initial impoundment of the Koyna reservoir and was associated with slip on a major pre-existing fault (Gupta, 2002). Subsequent seismicity displays a broader range of strike-slip and normal faulting mechanisms distributed across multiple sub-zones of the seismogenic system, indicating activation of subsidiary structures in response to evolving pore-pressure conditions and stress perturbations (Modak et al., 2022; Sutar et al., 2025). Together, these examples illustrate that reservoir-impoundment perturbations can be accommodated by fault systems of different structural scales and that secondary structures may play an important role in distributing deformation within tectonically heterogeneous fault networks.

The Castanhão observations fit this broader structural picture. The spatial organization of seismicity into distinct clusters associated with different parts of the fault system, together with the modeled reservoir-induced pore-pressure perturbations, suggests that the seismic response is strongly modulated by the pre-existing fault architecture. Reservoir loading may perturb the system through both poroelastic stress transfer and time-dependent pore-pressure diffusion, while the structural and hydraulic heterogeneity of the fault network controls where these perturbations are most effectively transmitted and where fault reactivation occurs.

Similar interactions between hydromechanical forcing and structurally heterogeneous fault systems have been proposed for Song Tranh 2 reservoir in central Vietnam (Orlecka-Sikora et al., 2023) and Reykjanes Peninsula in southwestern Iceland (Hrubcová et al., 2021). These conclusions, however, should be treated as speculative rather than definite, due to an incomplete seismological record.

This interpretation is also consistent with the broader concept of progressive rupture through heterogeneous fracture networks. Numerical simulations by Palgunadi et al. (2024) show that rupture in multiscale fracture systems can propagate through cascading activation of interacting fracture segments, with larger events emerging as progressively larger portions of

the network become involved. The spatial extent of activated fracture networks may approach the rupture dimensions associated with larger events, suggesting that the spatial extent of swarm activity could provide information on the potential magnitude of earthquakes that such fault systems are capable of hosting. In this context, the spatial extent of the activated fault system may be more informative about the potential scale of rupture than the magnitudes of the individual earthquakes recorded during the swarm. This is consistent with the fault-dimension estimates for Castanhão, which indicate a potential earthquake size substantially larger than the magnitudes observed in the available catalog.

3.4. Evolution of Seismic Strain Dynamics metrics

Given the progressive and spatially distributed seismic activation observed at the Castanhão Reservoir, we further examine whether its temporal evolution reveals systematic patterns in damage accumulation that may reflect the evolving state of the activated fault system. In case of artificial water reservoirs, the recent studies (Orlecka-Sikora et al., 2025a; Orlecka-Sikora et al., 2025b) indicate that the application of Seismic Strain Dynamics (*SSD*) metric – often, with a combination of seismic regularity coefficient (d_c) – can be used to define the stage of increased hazard of larger earthquakes. The *SSD* and d_c metrics were proposed to track the Earthquake Preparatory Process (EPP) and were tested for both natural and anthropogenic seismicity. They are particularly effective in the case of RTS (Orlecka-Sikora et al., 2025a, 2025b). To further complete this study, we calculate the *SSD* metric and seismic regularity coefficient d_c for the combined Castanhão Reservoir catalog (Figure 5). Despite the relatively small catalog size, that could be insufficient to support a robust quantitative analysis, the seismic regularity coefficient was evaluated and mimics the d_c behavior characteristic for RTS EPP.

The five-stage damage preparation model was previously proposed by Orlecka-Sikora et al. (2025a) and is defined by the temporal evolution of *SSD* metrics and d_c , which carry information on seismic strain variability. We investigated the temporal evolution of *SSD* metrics, quantifying the mean rate of cumulative seismic strain in a moving time window (to

assess the model's applicability, we use window $W = 20$). Mathematically, the SSD metric is defined by formula (7):

$$SSD(t) = \dot{\epsilon}(t) = \frac{\partial}{\partial t} \left(\sum_{i=1}^{n(t)} M_{0,i}^{\zeta}(t) \right) = \frac{\sum_{i=1}^n M_{0,i}^{\zeta}}{\Delta t_i}, \quad (7)$$

where $\dot{\epsilon}(t)$ represents strain rate, $\sum M_{0,i}^{\zeta}(t)$ is related to Accelerated Seismic Release (ASR) following a power law of time-to-failure, t is time, Δt_i is the inter-event times between successive earthquakes, $M_{0,i}$ expresses the i -th event's seismic moment, while n represents the number of events in the temporal window. Finally, SSD quantifies the Subcritical Fracture Growth (SFG) for $\zeta = 1/3$ (Main, 1999; Orlecka-Sikora and Cielesta, 2020). Such a choice of ζ is dictated by self-similar seismic event-source scaling, for which M_0 scales with the cube of a characteristic source dimension (Kanamori and Anderson, 1975). Letter i represents the i -th earthquake, while n stands for the number of events observed in a time window (Orlecka-Sikora et al., 2025a).

The event window selection is performed based on the size of the catalog. The small event window may impact the statistical robustness of the five-stage EPP interpretation. This, however, still lets us treat the results as indicative.

Similar assumptions ($W = 20$) were applied to computation of seismic regularity coefficient, which is given by formula (8). The coefficient d_c describes the parameters of occurrence of the earthquake related to its spatiotemporal variables (time t and hypocentral location— X , Y , Z), as well as its size (moment magnitude M_w) (Lasocki et al., 2024). Based on these parameters, the waiting time (dt) is calculated along with epicentral offsets (dR) (Lasocki et al., 2014). All of them – dt , dR , M_w – are normalized by transformation into equivalent dimensions (EDs) introduced by Lasocki (2014). The parameters are replaced with their cumulative distribution function (CDF) F : $F_{dt} = P(X \leq dt)$, $F_{dR} = P(Y \leq dR)$, and $F_{Mw} = P(Y \leq Mw)$.

$$d_c = \frac{2}{(n-1)n} \sum_{i=1}^{n-1} \sum_{j=i+1}^n \sqrt{(F_{dt,i} - F_{dt,j})^2 + (F_{dR,i} - F_{dR,j})^2 + (F_{Mw,i} - F_{Mw,j})^2}, \quad (8)$$

where n corresponds to the number of events within the moving time window. The seismic regularity coefficient quantifies the degree of spatiotemporal regularity in seismicity.

The EPP model distinguishes five stages (Figure 5): (1) Primary Stage (Stage I) – in which $\log(SSD)$ is relatively stable with a possible decrease. (2) Secondary Stage (Stage II) – $\log(SSD)$ continues to decrease. (3) Nucleation Stage (Stage III) – the transition from elastic to plastic behavior of the medium. Stage III marks the beginning of notable fracture growth. The $\log(SSD)$ decreases continuously until it reaches its minimum. (4) Hardening Stage (Stage IV) – the increase in deformation resistance of the medium. In this stage, the $\log(SSD)$ increases sharply to a high level, then decreases. (5) Failure Stage (Stage V) – This stage is marked by the approach to mechanical saturation, where the fault zone is extensively damaged but has not yet failed completely. The $\log(SSD)$ decreases to stabilize at a constant level.

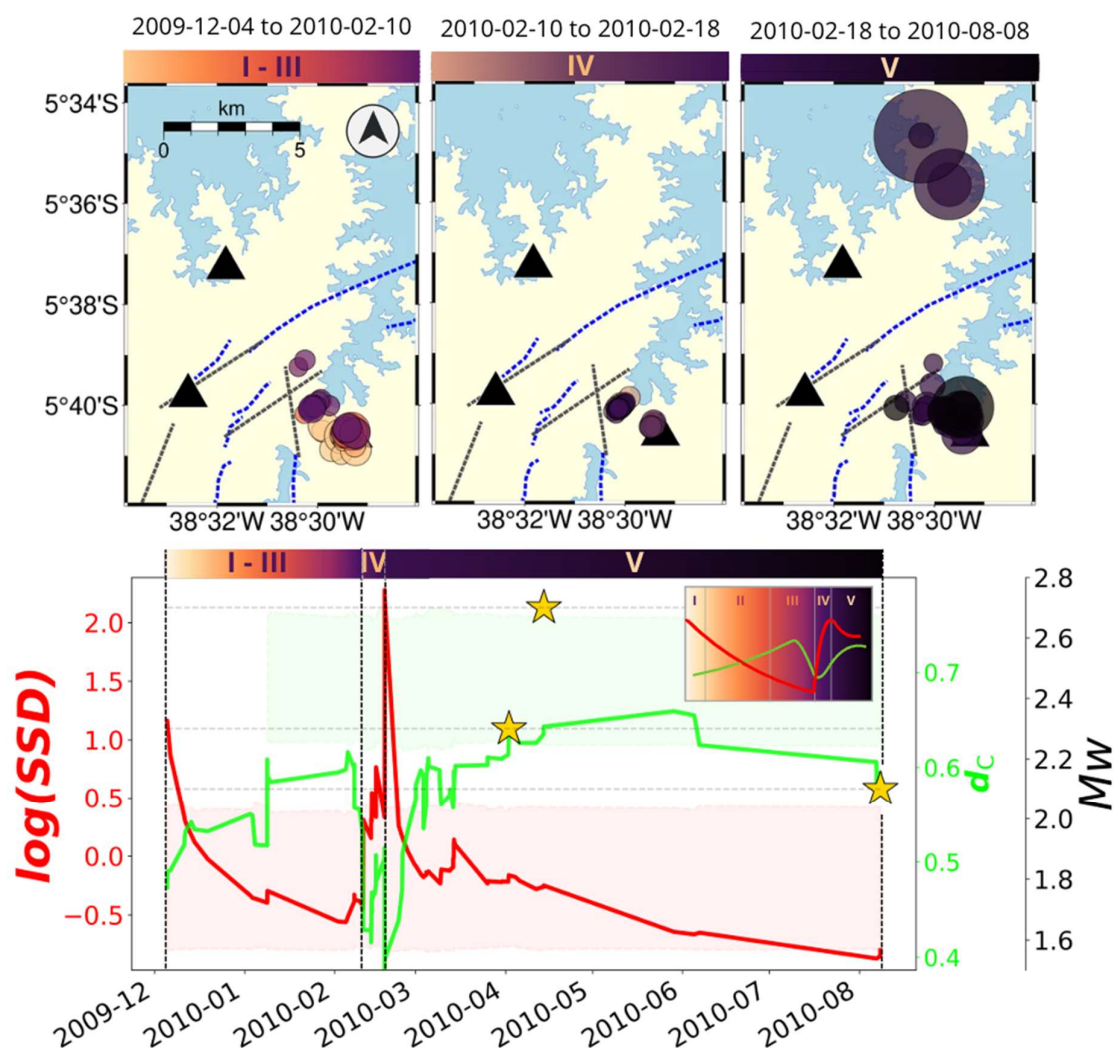


Figure 5. Temporal evolution of the log(SSD) metric and the seismic regularity coefficient d_c for seismic activity at the Castanhão Reservoir, calculated from the combined earthquake catalog and filtered for events with moment magnitudes between 0.4 and 2.7 ($M_c = 0.4$). A sliding window of $W = 20$ events was applied. Dashed black lines indicate the stages of the EPP, while golden stars mark the largest-magnitude earthquakes in the catalog together with their dates of occurrence. The red curve shows the evolution of log(SSD), and the green curve depicts the behavior of d_c . A schematic illustration of the expected log(SSD) and d_c patterns is shown in the right inset. The light red and green regions mark the 95% confidence intervals calculated for randomly ordered earthquakes without any clustering of

dt , dR , and/or M_w , respectively, for $\log(SSD)$ and d_c for a random earthquake process. The top panels show the maps of earthquakes occurring within specific EPP stages.

The progressive spatial reorganization of seismicity within the Castanhão fault system is accompanied by systematic changes in seismic strain dynamics. The joint evolution of SSD and the seismic regularity coefficient d_c shows a distinct temporal pattern during the period covered by the available catalog (Figure 5).

During Stages I–III, $\log(SSD)$ shows an overall decreasing trend, indicating a progressive reduction in the rate of seismic strain release. This trend is interrupted at the transition to Stage IV, when $\log(SSD)$ increases sharply during a short-lived episode of enhanced seismic strain release in February 2010. At approximately the same transition, d_c reaches its minimum and subsequently increases rapidly. The coincidence of the d_c minimum with the transient change in SSD marks a distinct transition in the evolution of the seismic process.

Stage V begins after the transient SSD increase and is characterized by a renewed, long-term decrease in $\log(SSD)$. In contrast, d_c continues to increase and subsequently remains at relatively high values. Thus, the later evolution of the sequence is characterized by the divergent evolution of the two parameters, decreasing SSD and elevated d_c , indicating a change in the organization of seismic deformation following the Stage IV transition. This temporal evolution is accompanied by a clear change in the spatial distribution of seismicity (Figure 5). During Stages I–III, the earthquakes are concentrated predominantly within the southern cluster. Stage IV remains spatially restricted to this part of the fault system, whereas Stage V is characterized by activity over a substantially broader area, including the northern cluster. The larger earthquakes of the analyzed sequence also occur during Stage V. The transition from Stage IV to Stage V therefore involves not only a change in SSD and d_c , but also a redistribution of seismic activity across different parts of the fault system.

Despite the limited catalog, d_c exhibits a characteristic evolution previously identified in other RTS sequences and laboratory stick-slip experiments (Orlecka-Sikora et al., 2025a, b). Its

decrease toward a minimum at the onset of Stage IV, followed by a rapid increase, occurs together with the characteristic *SSD* transition from the declining trajectory of Stages I–III to the transient Stage IV increase and subsequent Stage V decrease. The coupled evolution of *SSD* and d_c therefore provides complementary information on changes in the rate and organization of seismic deformation during the Castanhão sequence.

A closer comparison with the reservoir water-level evolution provides additional insight into the prolonged Stage V. Following the transition between Stage IV and V, *SSD* decreases over most of Stage V, whereas d_c recovers from its minimum and remains elevated, indicating that the system evolves within the post-localization regime despite the decreasing rate of seismic strain release. A later short-term increase in reservoir water level of approximately 0.5 m between March and April 2010 coincides with a secondary, much smaller *SSD* peak during Stage V. This correspondence indicates a transient increase in seismic strain release associated with a relatively small change in reservoir loading after the onset of localization. The two stronger earthquakes occur subsequently, on the long declining *SSD* trajectory. Thus, in the Castanhão sequence, the onset of shear localization precedes the stronger events by a substantial interval, and the localized deformation regime persists through a prolonged period of decreasing *SSD* and decreasing water level before these events occur. This apparent decoupling — stronger earthquakes occurring while both *SSD* and reservoir water level are declining — is consistent with the interpretation of d_c as a marker of a change in the deformation regime rather than of the instantaneous loading state. The sustained elevation of d_c after its minimum indicates that seismic strain release remains spatially concentrated on a reduced portion of the fault system, even as the system-averaged strain rate (*SSD*) falls. Under such a concentrated regime, the occurrence and magnitude of individual events are governed less by the current sign of the external forcing than by the stress state already established within the localized portion of the fault. This would explain why the two stronger earthquakes occur on a declining *SSD* and declining water-level trajectory: once localization has occurred, further evolution of the sequence appears to be controlled primarily by the

internal state of the fault system rather than by short-term reservoir loading trends. This separation between localization and the subsequent stronger events suggests that the d_c minimum identifies a change in the deformation regime rather than an immediate transition to dynamic failure. Subsequent reservoir-level variations may perturb strain release within this already localized system without necessarily producing immediate larger rupture.

4. Conclusions

This study examines the earthquake swarm that occurred in the Castanhão Reservoir area between late 2009 and 2010. Detailed waveform analysis and the application of the template-matching algorithm allowed us to construct an earthquake catalog of 227 visually verified seismic events. Moment magnitudes were successfully estimated for 219 events. To refine the catalog, double-difference relocation was performed to improve the spatiotemporal characterization of the swarm. This procedure enabled the relocation of 103 events. The initial Hypo71 catalog was enhanced using the HypoDD re-located events. The final product will be made available as a new Episode on the EPOS EPISODES Platform.

Analysis of the refined seismic catalog shows the earthquakes form two distinct clusters in map view and depth, likely corresponding to two fault strands or en-echelon segments within a secondary transtensional fault system. Diffuse seismicity occurs below the main rupture plane, suggesting off-fault damage or fluid-migration pathways.

The earthquakes associated with the Castanhão Reservoir appear to have been triggered by pore pressure changes induced by reservoir impoundment and fluid diffusion in the underlying rocks. These findings are consistent with results of other studies of RTS (do Nascimento et al., 2005; Orlecka-Sikora et al., 2023; Hrubcová et al., 2021) occurring in similar tectonic settings. These conclusions are also supported by previous studies in the Castanhão Reservoir region, which showed a correlation between seismic activity and variations of reservoir water levels in the early 2000s (Ferreira et al., 2008).

The EPP for the main event (M_w 2.1 on August 8th, 2010) can be tracked through the joint evolution of $\log(SSD)$ and d_c , consistent with the V-stage rock damage EPP model proposed

by Orlecka-Sikora et al. (2025a). Notably, the d_c minimum marking the onset of localization precedes the two strongest earthquakes of the sequence by a substantial interval, during which both *SSD* and reservoir water level decline. This indicates that localization marks a change in the deformation regime rather than an imminent transition to dynamic failure, and that the timing of subsequent larger events is not simply tied to the sign of the external loading trend.

Moreover, this study provides rare structural insight into the response of a secondary transtensional fault system to reservoir loading in an intraplate environment. The steep fault geometry and rupture dimensions are consistent with localized reactivation under moderate fluid overpressure. This supports the interpretation that this tectonic setting is structurally predisposed to RTS in. Overall, our study highlights the dynamic and predictive role of secondary transtensional tectonic settings in RTS. Their internal segmentation, stress concentration, and high permeability make such structures natural focal points for anthropogenic stress perturbations. This also has an important practical implication for seismic hazard assessment. Accordingly, infrastructure planning – involving fluid injection, extraction, or reservoir impoundment – should recognize mapped secondary transtensional fault systems as highly sensitive zones with an increased risk of swarm-type seismicity, cumulative strain release, and delayed but significant seismic hazard.

We acknowledge several limitations. The analysis would benefit from more complete seismological data; however, no pre-2009 records are available due to the network's operational period. A short monitoring period and a small number of stations limited the detection of small events that could occur before or during the network's deployment and operation. The catalogs and analyses presented here account for these limitations.

Data Availability Statement

The data (earthquake catalogs: the original Hypo71 catalog and the improved catalog including re-localized seismic events, both with KiwiTools-computed magnitudes) are available on IG PAS Data Portal: <https://geodata.igf.edu.pl/Castanhao/> (DOI:

https://doi.org/10.25171/InstGeoph_PAS_IGData_seismology-2026-018). These two catalogs are also planned to soon be made available on the Thematic Core Service: Anthropogenic Hazards EPISODES Platform as “Castanhão Episode”.

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Authors Contribution

HCi: original draft, writing, editing, conceptualization, data curation, formal analysis, investigation, methodology, software, visualization; BOS: writing, reviewing, formal analysis, methodology, software; ŁR: writing, reviewing, methodology; AFN: writing, reviewing, data curation; AV: writing, reviewing, software; JAF: reviewing, methodology, validation.

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