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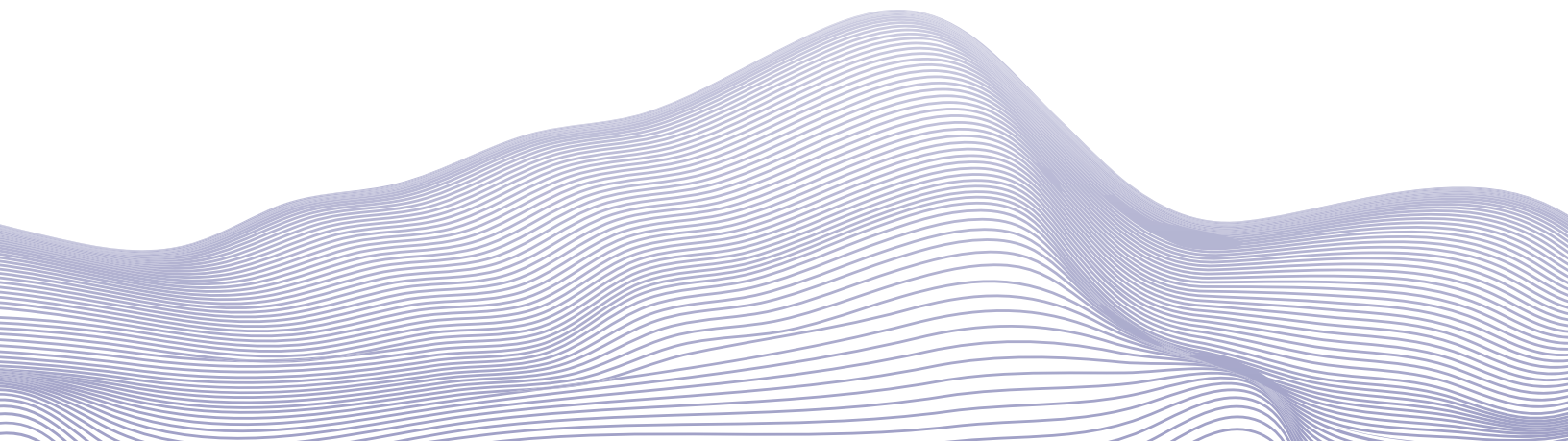
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RIESGO: A New Seismogeodetic Network Spanning the States of Guerrero and Oaxaca in The Mexican Subduction Zone

Gerardo Suárez^{1,a} , José Antonio Santiago-Santiago^{1,b} , Diego Espinosa^{1,c} ,
Ekaterina Kazachkina^{1,d} , Sergio Aguilar-Flores^{2,e}  and Ketzallina Flores-Ibarra^{3,f} 

Abstract

A seismogeodetic network of twenty stations has been installed along the subduction zone of the states of Guerrero and Oaxaca (RIESGO). The purpose of the network is to continuously monitor the tectonic deformation of the coast along the Mexican subduction zone. Also, the mission of the seismogeodetic network is to estimate the magnitude of large and great earthquakes based on the length of the fault ruptured. The network was built in two stages. The first fifteen instruments were installed in the state of Guerrero and straddle the Guerrero seismic gap. This segment of the subduction zone has not experienced earthquakes of magnitude $M_w > 7$ since the turn of the 20th century and has been the source of intense study for many years. The second phase of RIESGO was installed in the state of Oaxaca, where five instruments cover a large fraction of the subduction zone here. RIESGO outputs an acceleration stream sampled at 200 Hz. A high precision geodetic stream is produced at two samples per second, as a result of the application of an RTX correction and of a Kalman filter that uses the accelerometer data as ground truth to correct the geodetic signals in real time. A comparison between the high-precision geodetic data in real time and the post-processed data using the program GAMIT shows practically identical results. These results support the use of RIESGO in real time monitoring of the subduction zone. To date, the only large earthquake in the Mexican subduction zone recorded by RIESGO is the 8 September 2021 Acapulco earthquake (M_w 7.1). This event was recorded when the seismogeodetic network was in the initial stages. Nevertheless, it was well recorded by the station located in the city of Acapulco, onshore of the epicenter. The recorded displacement seismogram shows a co-seismic deformation of ~20 cm, similar to the results obtained later by other authors using satellite interferometry (Interferometric Synthetic Aperture Radar InSAR), modeling of the source, and data from a tide gauge station. These initial results show that the seismogeodetic network is a robust and reliable tool to monitor the subduction zone in real time and in the long term.

Key words: Seismogeodesy, subduction, Middle American Trench, Guerrero gap, Oaxaca.

Resumen

Una red sismogeodésica de veinte estaciones fue instalada a lo largo de la zona de subducción de los estados de Guerrero y Oaxaca (Red Sismogeodésica de Guerrero y Oaxaca, RIESGO). El objetivo de la red es monitorear de forma continua la deformación tectónica de esta parte de la zona de subducción de México. Asimismo, la red sismogeodésica tiene como misión estimar la magnitud de grandes terremotos, con base en la longitud de la ruptura. La red se construyó en dos etapas. Los primeros quince instrumentos se instalaron en el estado de Guerrero y abarcan la brecha sísmica de Guerrero. Este segmento de la zona de subducción no ha experimentado terremotos de magnitud $M_w > 7$ desde principios del siglo XX y ha sido objeto de intenso estudio durante muchos años. La segunda fase de RIESGO se instaló en el estado de Oaxaca, donde cinco instrumentos cubren una gran parte de la zona de subducción. RIESGO genera un flujo de datos de aceleración muestreados a 200 Hz. También, genera datos geodésicos de alta precisión muestreados a 2 Hz, como resultado de la aplicación de una corrección RTX y un filtro de Kalman, que utiliza los datos del acelerómetro como referencia para corregir las señales geodésicas en tiempo real. Una comparación entre los datos geodésicos de alta precisión en tiempo real y los datos procesados posteriormente con el programa GAMIT, muestra resultados prácticamente idénticos. Estos resultados respaldan la utilidad de RIESGO para el monitoreo en tiempo real de la zona de subducción. Hasta la fecha, el único gran terremoto en la zona de subducción mexicana que ha sido registrado por RIESGO es el terremoto de Acapulco del 8 de septiembre de 2021 (M_w 7.1). Este evento se registró cuando la red sismogeodésica se encontraba en sus etapas iniciales. Sin embargo, el sismo fue registrado por la estación ubicada en la ciudad de Acapulco, cerca del epicentro. El sismograma de desplazamiento registrado muestra una deformación cosísmica de aproximadamente 20 cm, similar a los resultados obtenidos posteriormente por otros autores mediante interferometría satelital (Interferometric Synthetic Aperture Radar, InSAR), modelado de la fuente y con base en un registro mareográfico. Estos resultados iniciales demuestran que la red sismogeodésica es una herramienta robusta y fiable para monitorizar la zona de subducción en tiempo real y a largo plazo.

Palabras clave: Sismogeodesia, Subducción, Trinchera Mesoamericana, Brecha de Guerrero, Oaxaca.

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1. Introduction

The Guerrero gap has attracted attention for many years, as no major earthquakes ($M_w > 7.5$) have occurred along the subduction zone since the early 20th century (Figure 2) (Kelleher *et al.*, 1973; Singh *et al.*, 1981; Suárez *et al.*, 1990; Kostoglodov *et al.*, 2003; Radiguet *et al.*, 2012; Plata-Martínez *et al.*, 2021; Cruz-Atienza *et al.*, 2025). Hence, this segment of the Mexican subduction zone is expected to produce eventually a major earthquake. The seismic energy accumulated in the Guerrero gap is estimated to be large enough to produce potential earthquakes of magnitude between M_w 8.1 to 8.4 (e.g., Singh and Mortera, 1991; Valdés-González and Novelo-Casanova, 1998) (Figure 2).

In order to monitor this segment of the subduction zone, presumed to be of high seismic potential, a seismogeodetic network (Red Sismogeodésica de los Estados de Guerrero y Oaxaca (RIESGO) was installed to continually monitor the seismicity and the short and long-term deformation of the Mexican subduction zone. From a civil protection perspective, there are two key objectives in the design of RIESGO. The first one is to estimate the size of very large earthquakes, larger than $M_w \sim 7.5$, for which the seismic instrumentation of the Seismological Service of Mexico (SSN, Servicio Sismológico Nacional, in Spanish) typically saturates. The advantage of seismogeodetic data over broad-band seismographs in the estimation of magnitude of large and great earthquakes has been extensively demonstrated (e.g., Crowell *et al.*, 2013; Melgar *et al.*, 2013; Chen *et al.*, 2018).

The concept behind this is that, based on the high-quality and high frequency seismogeodetic signal, the deformation of the subduction zone along the coast can be recovered in real time. The estimated fault length based on the deformation of the coast is used as the key parameter to estimate magnitude (e.g., Blewitt *et al.*, 2006). The fault length can be calculated in real time, so magnitude can be estimated within a few seconds after the origin time. Otherwise, the magnitude estimate has to wait for several minutes, until teleseismic data is retrieved. In addition, the length of the fault may be used to estimate areas of potential damage in communities along the coast and neighboring regions.

The second objective, from a civil protection outlook, is to estimate the size of a potential tsunami produced by a major subduction earthquake. Using the deformation of the coast that is observed in real-time by RIESGO, the displacement on the fault defining the interplate contact may be estimated. This information is used to calculate the deformation of the ocean floor and, in turn, calculate the areal extent and inundation height of the potential tsunami. This project is under development and the results would be provided to the Center for Tsunami Alerts, which is under the responsibility of the SEMAR (CAT, Centro

de Alerta de Tsunamis, in Spanish). The network is relatively low-cost as it operates automatically and the communication links are provided free of charge by the Mexican Navy (SEMAR).

From a scientific perspective, the network was designed to provide continuous monitoring of the deformation of the coastline, on this portion of the Mexican subduction zone. The geometry and maximum seismogenic depth of the interplate contact in the Middle American trench shows that the largest vertical deformation during subduction events takes place near the coastline (Suárez and Sánchez, 1996). Thus, to capture the pre-seismic, co-seismic, and post-seismic deformation of the coast in real time, RIESGO was built on rocky outcrops as close to the coast as possible. This requirement presented a challenge to find the better suited locations that meet the following criteria: hard rock siting, vicinity to the coastline, sky view without obstruction, and the absence of radio frequency interferences and of reflective surfaces, to avoid signal multipathing. The location of the sites was also designed to complement the existing geodetic stations installed by other projects and the data is being used jointly.

2. Network Description

RIESGO is installed along the coast of Guerrero and Oaxaca in southern Mexico. It consists of twenty stations that have an accelerometer and a GNSS receiver (Figure 1). The accelerometer data is used as ground truth to correct errors in the GNSS receivers using a recursive Kalman filter. As explained below in more detail, the Kalman filter estimates a correction of errors present in the geodetic data to produce a real-time displacement time signal with a 2 Hz frequency. RIESGO was built in two stages. The first phase was designed to monitor the Guerrero seismic gap with fifteen instruments.

In the second phase, RIESGO was extended to the southeast to include the state of Oaxaca, using five additional stations to monitor this part of the subduction zone (Figure 1). At the time of writing this manuscript, permits for two of the five new sites in Oaxaca are still under negotiation; the installation of the twenty stations is expected to conclude in early 2026. The seismogeodetic stations in Oaxaca cover the Ometepe segment of the subduction zone, located at the border of the states of Guerrero and Oaxaca (Figures 1 and 2). This region has been identified as a region of high seismic potential (Singh *et al.*, 1981; Nishenko and Singh, 1987; Mendoza and Martínez-López, 2022; Martínez-López, 2023; Yan *et al.*, 2022). This segment of the subduction zone near Ometepe region, exhibits an unusually high rate of seismicity, which has been attributed to presence of fluids in the subduction zone (Legrand *et al.*, 2021).

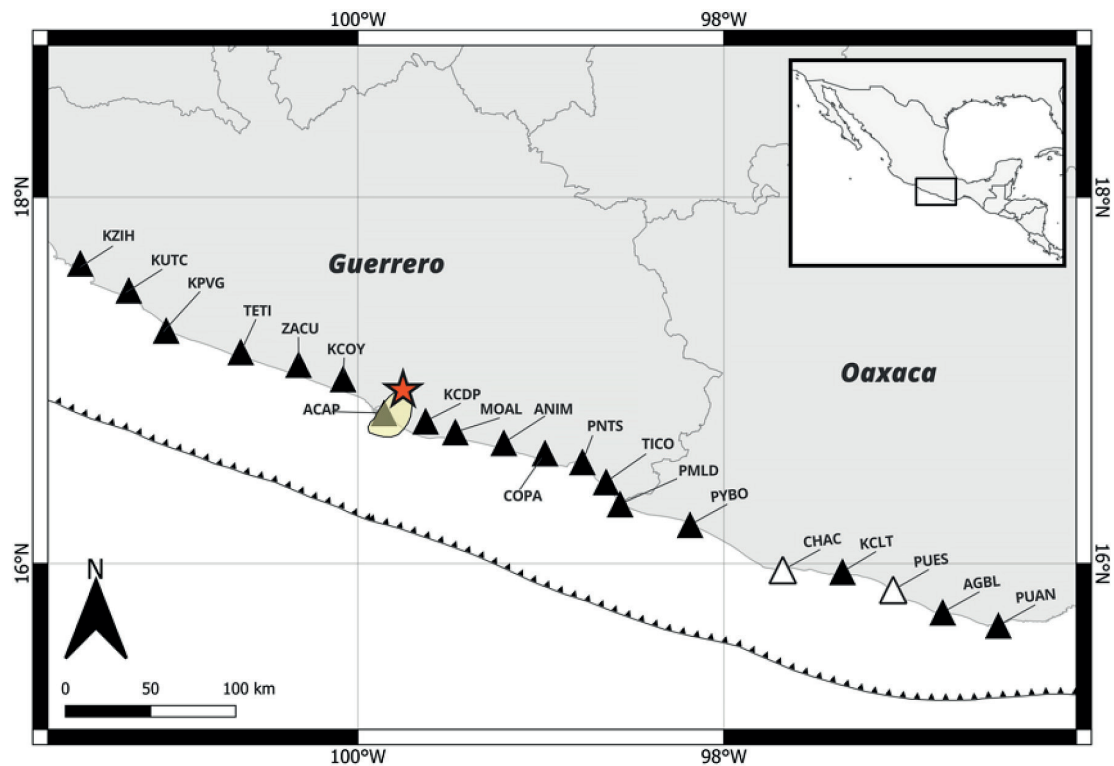


Figure 1. Location of the twenty stations of the Seismogeodetic Network of Guerrero and Oaxaca. The black triangles are operational stations and the two white triangles are the sites that are being built. The rupture zone of the 8 September 2021 Acapulco earthquake is shown as a yellow area. The red star is the epicenter.

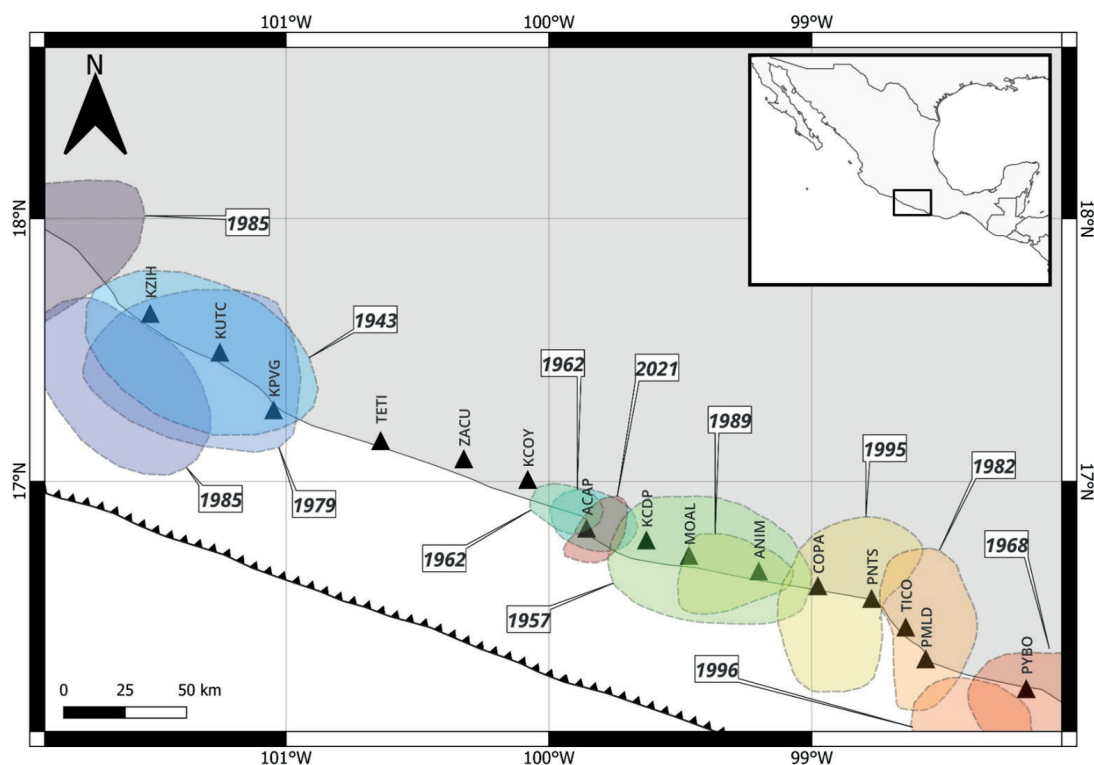


Figure 2. Stations of the Seismogeodetic Network of Guerrero-Oaxaca (triangles) shown together with the rupture areas of the major earthquakes in the region (shaded areas). The dentate line represents the Middle American trench. The Ometepec area is where the 1982 earthquake took place, near the PMLD station.

Two earthquakes of moderate magnitude (M_w 6.9, 7.0) took place in Ometepec within five hours of each other on 7 June 1982 (Astiz and Kanamori, 1984) (Figure 2). More recently, an earthquake took place in the same region on 20 March 2012 (M_w 7.5) (Universidad Nacional Autónoma de México, Seismology Group, 2013). The Oaxaca subduction zone exhibits a complex behavior of seismic ruptures of variable length. Quasi-repeater earthquakes have occurred in recent years where a sequence of four earthquakes took place in 1928 (Singh *et al.*, 2024; Garza-Girón *et al.*, 2025). The region where the event of October 1928 occurred is the only one where no large earthquakes have taken place since, indicating an area that is potentially seismically locked (Garza-Girón *et al.*, 2025; Martínez-López *et al.*, 2025). No great earthquakes have occurred in the last 96 years, suggesting high accumulation of elastic strain that may generate potentially an earthquake of magnitude M_w 7.8.

A great earthquake, M_w 8.6, ruptured this segment of the Oaxaca subduction zone in 1787. (Suárez and Albiní, 2009; Suárez *et al.*, 2020). This event is the greatest earthquake recorded in the Middle American subduction zone and underlines the variable rupture behavior of under thrusting, interplate events in this region. This seismically complex region in the Oaxaca subduction zone is now covered by seismogeodetic stations of RIESGO in Oaxaca and complements the other seismic and geodetic instruments installed by other agencies.

3. Instrumentation and Data Flow

3.1. Instruments Used in RIESGO

The Trimble Kestrel SeismoGeodetic System, SG160-09 was the instrument selected for RIESGO (Passmore *et al.*, 2014). The instrument combines seismic recording of a forced-balance triaxial accelerometer with a 24-bit A/D converter, with a 336-channel GNSS receiver, which tracks the constellations GPS, Galileo, GLONASS, BeiDou, and QZSS. The receiver incorporates GNSS Precise Point Positioning (PPP) using Trimble RTX technology with satellite clock and orbit corrections delivered in real time over the L-Band signal. The Kestrel system combines the GNSS and the acceleration measurements using the Kalman filter algorithm to create a high rate (2 Hz), real-time displacement with sufficient accuracy and very low latency for earthquake analysis and crustal deformation studies.

The accelerometer data is sampled continuously at 200 Hz. The data stream from the GNSS receiver is corrected using the RTX (Real-Time eXtended) protocol in real time. The RTX protocol delivers real-time, high-accuracy satellite ephemerides to GNSS stations. The correction is a high-precision positioning technique

developed by Trimble Inc. that improves raw satellite positioning without requiring a nearby base station. Instead of relying on local differential corrections (like RTK), RTX uses a global network of reference stations and precise satellite orbit and clock models delivered via satellite. The RTX protocol corrects the variations in orbital ephemerides of the satellites, using the data collected by a global network of ground tracking stations, and delivering high-accuracy satellite corrections to the GNSS receivers in real time via satellite, directly to the antenna of the Kestrel instruments.

The GNSS data, that includes the RTX correction, is passed through a Kalman filter, which uses the accelerometer data in real time as the ground truth for displacement of the ground. A Kalman filter is a recursive algorithm used to estimate the position determined by the GNSS instruments. It works in two repeating steps: a prediction, where it uses the accelerometric data to forecast the next state, and an update, where it corrects the position predicted by the GNSS data using these new observations and weighting them according to their uncertainties. Mathematically, it assumes Gaussian noise, producing an optimal estimate in the least-squares sense.

The result is a real-time, data stream of high precision ground motion data, which is transmitted via the satellite links to the central station in Mexico City. The corrected geodetic data is delivered in a 2 Hz data stream (Figure 3). The Kestrel also generates three channels of error estimates. In order to have the capability for post processing the raw GNSS data using different programs, the instruments upload the raw data at the end of the day to an FTP server to the central facility in Mexico City. This allows to process the raw data *a posteriori* to potential users using the program of their choice. Below, we show the comparison between the post-processed data using the program *GAMIT-GLOBK 10.71* and the output of the Kalman filter.

GAMIT is part of the *GAMIT/GLOBK* package developed at the Massachusetts Institute of Technology (MIT). This software produces high-precision GNSS data. It uses the corrections of satellite orbits and clocks, atmospheric delays, and tropospheric and ionospheric effects determined by various agencies several weeks after the data is collected. Based on these corrections, *GAMIT* performs a least-squares adjustment of the data.

3.2. Communications

The locations necessary for this type of network are sites as close to the coastline as possible and anchored on rock. In most cases, the locations selected are relatively isolated and lack internet and telephone connectivity. Also, in most cases, there is no electricity. Therefore, an autonomous communications system was needed to provide reliable real-time and continuous communications based on low power transmitters.

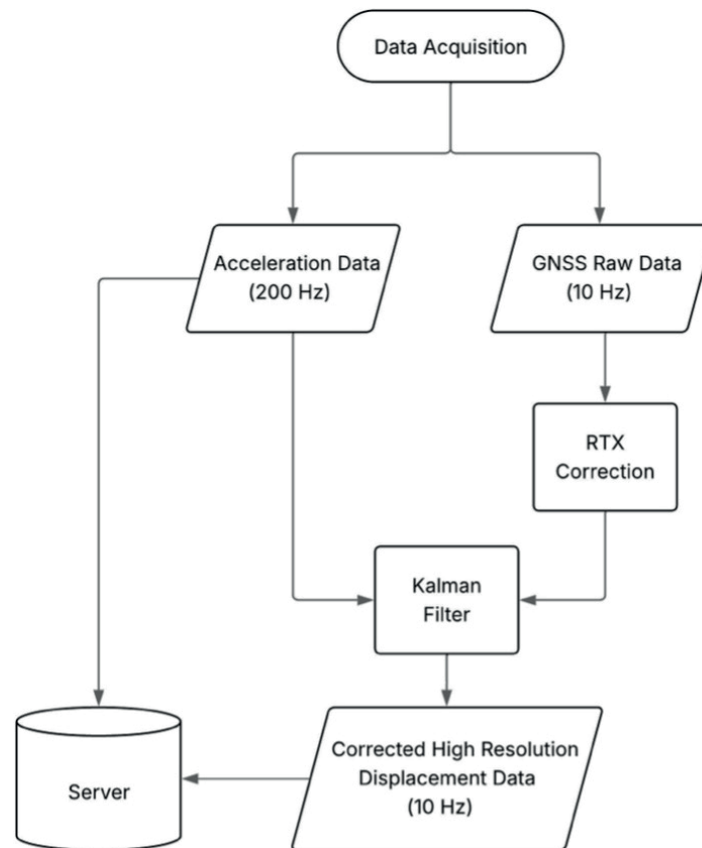


Figure 3. Data flow of the Kestrel seismogeodetic instrument. The acceleration data is sampled at 200 Hz. The incoming GNSS stream is corrected by the RTX protocol and passed through the Kalman filter to obtain a stream of high-quality, real time geodetic data.

A satellite internet terminal called UNI 3500, manufactured by Hughes Communications, was selected. It is a rugged and compact unit with very low power consumption, capable of sending and receiving IP packets via Ethernet. During the tests conducted in the SSN, the UNI 3500 terminals proved to be a robust and reliable communications link in real time. The sampling of the geodetic signals and of the error channels was reduced from the default 10 Hz to 2 Hz. This sampling rate is sufficient for the scientific purposes of this project and reduces the bandwidth demand on the satellite platform.

The satellite used by the UNI 3500 system is called MEXSAT. MEXSAT provides the communications link on an L-Band platform that is of exclusive use for the military and security services of Mexico. The Mexican Navy, considering that this project contributes to their responsibilities of providing timely tsunami warnings and that it improves the capability of estimating the magnitude of great earthquakes in the subduction zone, generously allowed the use of their platform on the MEXSAT satellite. These satellite internet links provide reliable and effective communication to the central facility at the Instituto de Geofísica, UNAM, in Mexico City. Figure

4 shows a photograph of a typical RIESGO station and its different components.

3.3. Central Recording Facility

The data from the seismogeodetic network are transmitted via the MEXSAT satellite and downloaded at the hub of the national institution that operates the Mexican Satellite System. From there, the data are sent to the Instituto de Geofísica, UNAM via a dedicated internet connection. At UNAM, the data are received in real time by the main server, which stores them on hard disk, and sends it to a second server (mirror). All data are ultimately stored on a high-capacity storage server, allowing access to potential users.

The second server is continuously performing state-of-health analyses using software produced in house, to evaluate the performance of RIESGO. As an example, a helicorder-type record is refreshed every five minutes to monitor gaps in the data from all stations (Figure 5). The monitoring and trouble-shooting tools also measure real-time and long-term voltage variations, data availability, the number of operating stations of the network.



Figure 4. Seismogeodetic station *Playa Banco de Oro* (PYBO) in the state of Oaxaca (Figure 1). On the vault cover is the solar panel that independently feeds the station. On top of the concrete column the GNSS antenna is protected by a *radome* and on the side of the column is the UNI 3500 L-Band internet terminal (left). The *radome*, short for radar dome, is a protective enclosure that protects the GNSS antenna while allowing electromagnetic signals (radio, GNSS, radar) to pass through with minimal distortion. The concrete vault hosts the Kestrel seismogeodetic instruments, the ethernet switch, the batteries, and the solar charge controller (right).

4. Evaluation of the Data Quality

One of the key questions was to explore the accuracy of the seismogeodetic data obtained in real time by the Kestrel instruments. To this end, the real time data were compared with the post processed results using the GAMIT software. As mentioned before, to process the raw geodetic data using GAMIT, one has to wait for several days for the final version of the corrections to become available.

Here, we show the comparison of the two data streams, real time locations from the seismogeodetic instruments and the data processed with GAMIT (Figures 6 and 7). Figure 6 shows this comparison for station KUTC, a station in the state of Guerrero, in the rupture area of the 21 September 1985 aftershock of the great Michoacan earthquake of 19 September (see Figures 1 and 2). The comparison clearly shows that the daily averages of the real-time geodetic data are practically identical to the post-processed GNSS stream using the *GAMIT* software package. The geodetic displacements obtained in real time show an accuracy of ~ 1 cm, similar to the post-processed data. Interestingly, the scatter in the vertical component is smaller in the real time data obtained by the Kestrel, when compared with the data processed several days after, using the GAMIT software.

A similar result is shown for the station ACAP, located in the city of Acapulco (Figure 7). Serendipitously, this station was installed a few weeks before the occurrence of the 8 September 2021 earthquake (M_w 7.1) in Acapulco (7 September in local time). The earthquake was recorded by the Acapulco seismogeodetic station, as we show below. The 2021 earthquake is an important event in the context of the Guerrero seismic gap. This is the first earthquake in this segment of the subduction zone since the two large earthquakes (M_w 6.7 and 7.0) that took place in 1962 (Ortiz *et al.*, 2000). Some authors interpret the Acapulco earthquake as a possible preamble to the reactivation of this segment of the subduction zone (e.g., Mendoza and Martínez-López, 2021; Iglesias *et al.*, 2022).

Of note is that the post-seismic deformation of the subduction zone is almost identical in both, the real-time output and the data processed using GAMIT. Post-seismic slip of the Acapulco earthquake lasting almost a year after the occurrence of the earthquake is clearly visible in the time series. Again, the results show that real-time data is as accurate and robust as data analyzed with GAMIT afterwards, proving that the locations obtained in real time by the seismogeodetic instruments. These results confirm that the RIESGO may be used for real time monitoring of the seismic behavior of the subduction zone in southern Mexico.

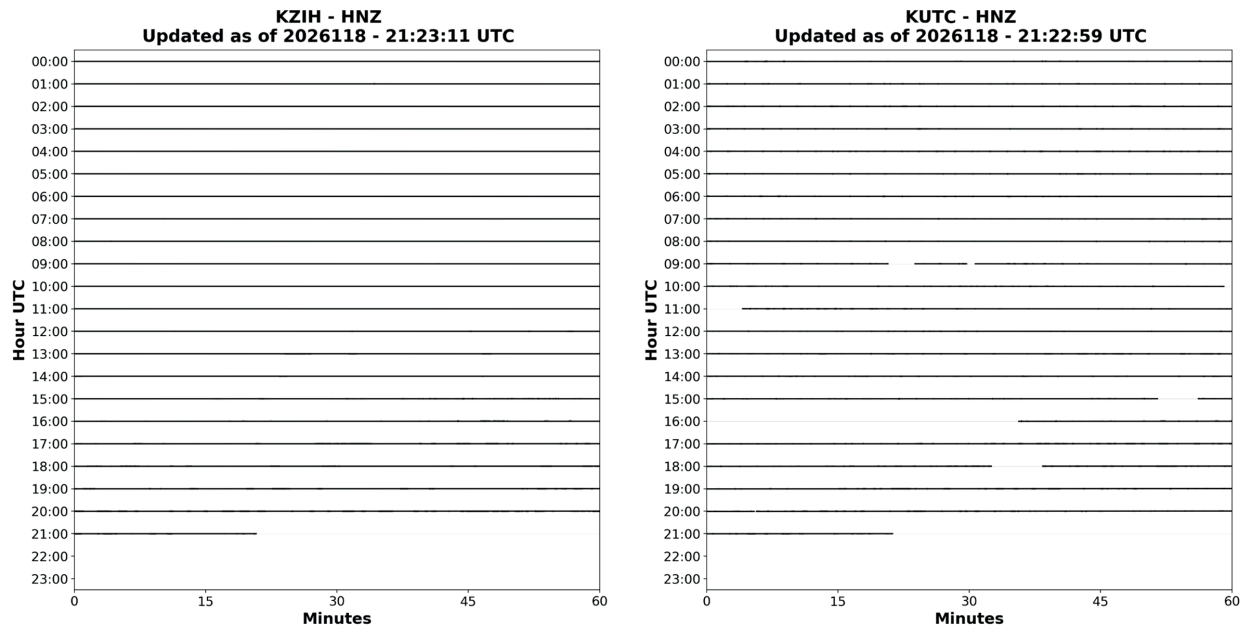


Figure 5. Example of the virtual helicorder records of two stations, KZIH and KUTC (Figure 1). The horizontal lines are one hour long. The graph shows data arrival in real time. As an example, the gaps in station KUTC reflect delays in data arrival. The graph is refreshed every five minutes. The monitoring tools depict the data continuously and for all stations.

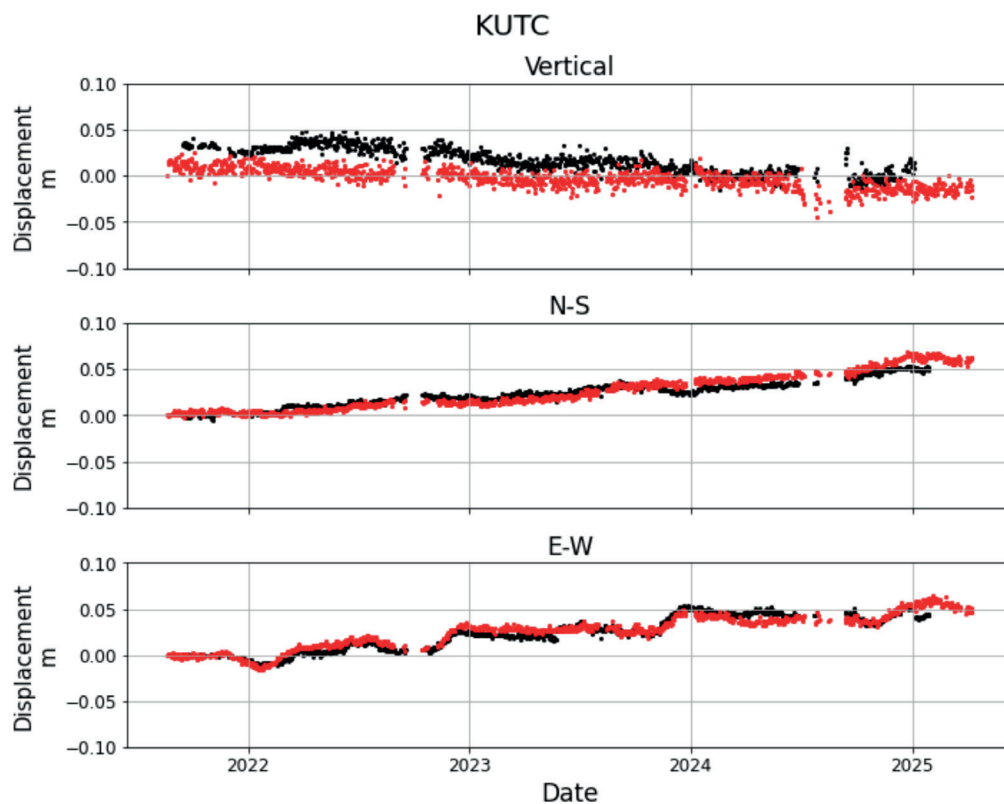


Figure 6. Comparison at station KUTC in northern Guerrero state, between the real time geodetic locations estimated in real time by the seismogeodetic instruments, by applying the RTX correction and the Kalman filter (black dots) and the daily averages resulting by processing the data using GAMIT several days after (red dots).

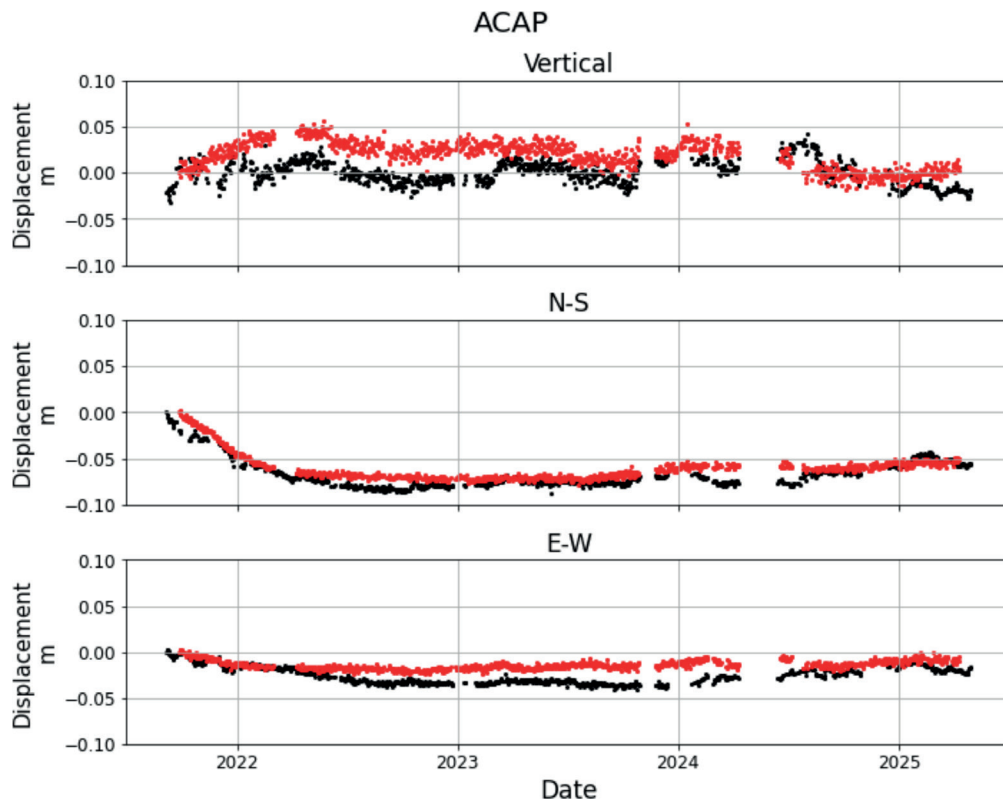


Figure 7. Comparison at station Acapulco (ACAP) between the real time geodetic locations estimated by the seismogeodetic instruments, by applying the RTX correction and the Kalman filter (black dots), and the daily averages resulting by processing the data using GAMIT (red dots), several days after. Notice the post-seismic slip after the 8 September 2021, Acapulco earthquake.

It is illustrating to see the residuals between these two data streams: the real time output of the seismogeodetic instrument and the post processed data using GAMIT. The residuals are defined as the difference between these two observations. These differences are generally less than ~ 1.5 cm, stressing the value of the seismogeodetic network to monitor the deformation of the subduction zone in real time (Figure 8).

5. Displacement Seismogram of the 8 September 2021 Earthquake

On 8 September 2021 an earthquake magnitude M_w 7.1 took place in the Middle American subduction zone. The epicenter of the earthquake was located in the city of Acapulco (Figures 1 and 2). ACAP, the seismogeodetic station in Acapulco, was installed a few weeks before the earthquake. It was one of the first stations of the network to be fully operational. The location of the station is very close to the shoreline, on the granitic outcrops surrounding the city (Figures 1 and 2). The earthquake of 8 September provided the first displacement seismogram to be recorded by RIESGO (Figures 9a, b).

The Acapulco earthquake ruptured the interplate contact beneath the city of Acapulco (Figure 2). The displacement record of the Acapulco earthquake on the seismogeodetic station Acapulco (ACAP) shows a co-seismic deformation of ~ 20 cm in the vertical component (Figure 9a). A co-seismic deformation of ~ 15 cm of to the west also reflects the tectonic rebound of the upper plate; practically no deformation on is observed in the N-S component. The co-seismic deformation of the coast inferred from the InSAR, (Interferometric Synthetic Aperture Radar), shows positive deformation in the Line of Sight of the satellite (LoS) of ~ 15 to 20 cm, reflecting the uplift of the coast (Melgar *et al.*, 2022; Cruz-Atienza *et al.*, 2025). A tide gauge meter in Acapulco also shows an uplift of the coast of ~ 16 cm (Arroyo *et al.*, 2025; Melgar *et al.*, 2022). Thus, the deformation reflected by the ACAP seismogeodetic instrument agrees with the other measurements of co-seismic coastal uplift.

The co-seismic static displacement of the upper continental plate was also observed in the two neighboring stations, KCOY y KCDP (Figure 2). However, the observed static deformation of the continental plate at these two stations is of only a few centimeters, as both lie on the edges of the mapped rupture

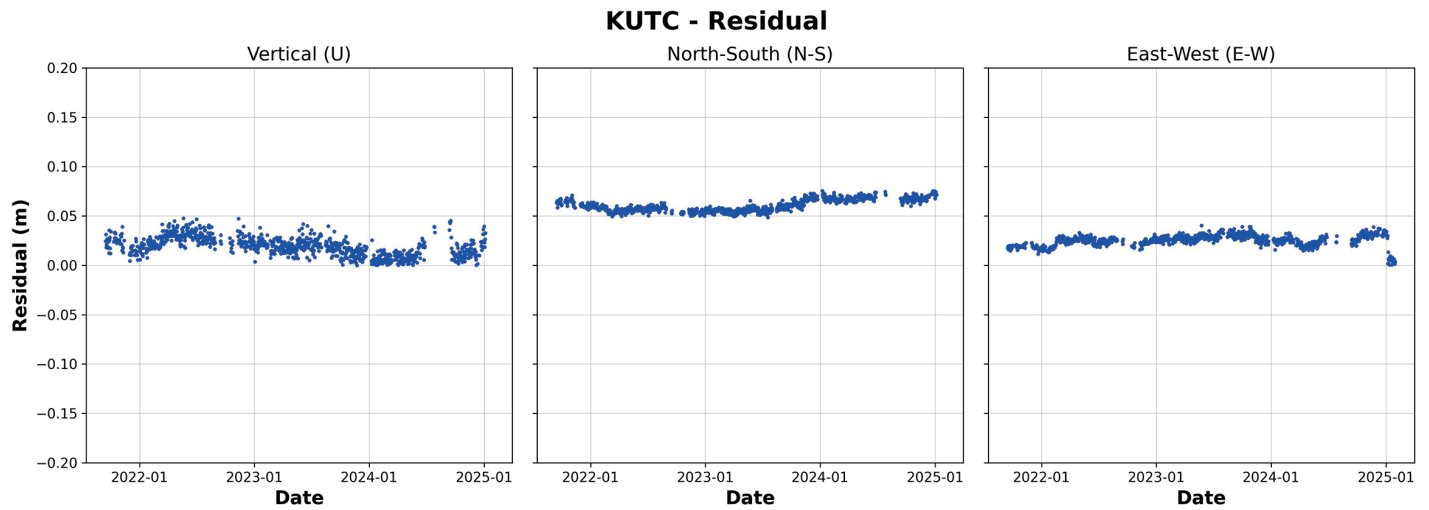


Figure 8. Difference between the real time displacement obtained in real time by the seismogeodetic station KUTC (Figure 1) and the post processed data using GAMIT. This example shows that the difference between these two datasets is less than ~ 1.5 cm.

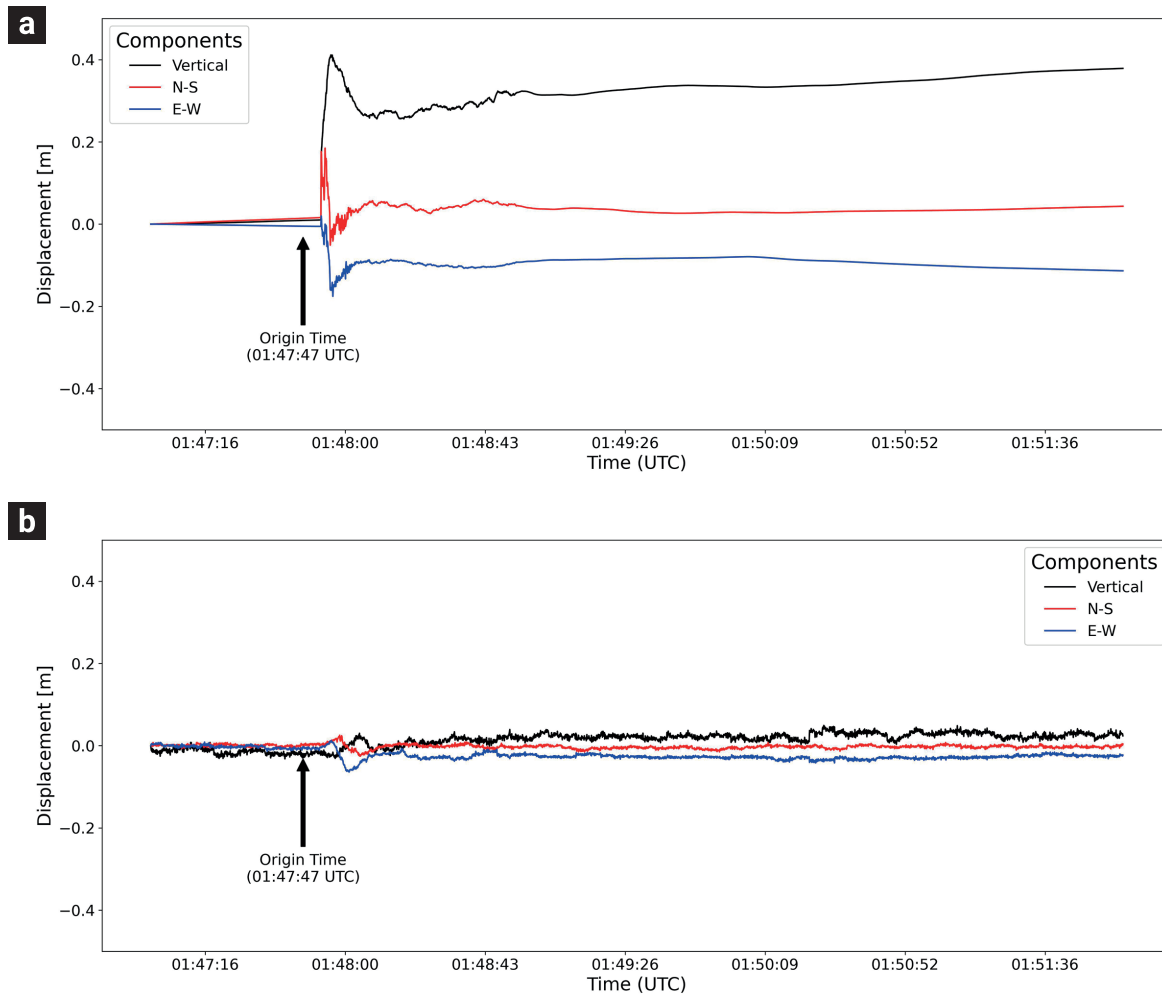


Figure 9. a) Displacement seismogram of the 8 September 2021 Acapulco earthquake ($M_w 7.1$) at station ACAP. The vertical black arrow indicates the origin time of the event, and the three-components of the displacement seismograms are shown as: vertical (black), north-south (blue), and east-west (red); b) Displacement seismogram of the Acapulco earthquake, recorded at station Coyoaca (KCOY) (see Figure 1). The co-seismic displacement is within the error of the measurements. However, the P and S waves were recorded by the station.

(Melgar *et al.*, 2022; Cruz-Atienza *et al.*, 2025). Station KCOY (Coyuca, Guerrero) is to the north of the rupture area of the 2021 Acapulco earthquake. A co-seismic deformation of ~5 cm is observed in KCOY (Figure 9b). The P and S waves were recorded by this seismogeodetic station (Figure 9b). Thus, these first displacement seismograms demonstrate the capacity of the seismogeodetic network to monitor the deformation of the coastline within a few centimeters.

6. Conclusions

RIESGO has operated continuously since the first stations were installed in 2021. Despite some data gaps due to the learning curve of the communication modems, the data reflect the continuous deformation of the upper continental plate due to the subduction of the Cocos plate beneath it as evidenced, for example, in station KUTC (Figure 6).

Since the installation of the RIESGO, several projects have emerged. An algorithm was developed that estimates the magnitude of large and great earthquakes using the length of the rupture zone, determined from the deformation of the coastline (Espinosa-Kuswara, 2025). In this case, we follow the methodology proposed by Singh *et al.*, (2012). Also, important progress has been made in using the RIESGO to monitor the deformation of the coastline in the subduction zone due to the interaction between the Cocos and North American plates.

Finally, we developed a set of templates of potential large and great earthquakes in the subduction zone, for which deformation of the ocean floor has been mapped in grids. Hypothetical tsunamis produced by earthquakes with ruptures similar to those predicted will be used to estimate tsunami heights and coastal areas likely affected by inundation. The results of these tasks are being prepared for publication.

7. Conflict of Interest

The authors have no conflict of interest in any of the topics included in this manuscript.

8. Acknowledgements

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9. Data Availability

The data is available upon request to the principal author or through the Mexican Seismological Network.

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Michoacán Seismic and Tremor (MIST) Network: Perspectives, preliminary results and future directions to understand the role of the Cocos plate and the monogenetic volcanism

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Abstract

The Michoacan Seismic and Tremor (MIST) network is a follow-up project of the Middle America Subduction Experiment (MASE) and the Veracruz-Oaxaca Network (VEOX). These former experiments consisted of linear arrays perpendicular to the trench that generated large waveform datasets to characterize the subduction of the Cocos plate better. While the MASE array focused mainly on the flat segment of the subduction zone in the state of Guerrero, Mexico the VEOX array examined the changes in geometry towards a steep subduction in the Tehuantepec Isthmus. The MIST network is being deployed in Michoacan state in western Mexico in an area where volcanic and seismic processes coexist. Besides the joint goal of characterizing the shape and rheology of the subducting slab, the MIST array will also explore the seismic swarms likely due to the magmatic process that forms the monogenetic Michoacan-Guanajuato volcanic field. Here, we review previous studies to gain an understanding of the subduction of the Cocos plate, as well as the current knowledge regarding seismic and volcanic activity. Additionally, we present the results of applying H/V directional ratios at low frequencies and how they may be used to infer the dip and strike of the subducting slab. We also report tremor detections along the array and what the relationship with both the subduction process and volcanic processes might be. Finally, we discuss our plans to expand the network and how we may continue to develop further analysis.

Key words: Michoacan, Mexico, H/V directional, temporary networks, tremor, monogenetic volcanism, subduction.

Resumen

La Red Sísmica y de Tremores de Michoacán (MIST, por sus siglas en inglés) es un experimento que da continuidad a las redes sísmicas temporales MASE (Middle America Subduction Experiment) y VEOX (Veracruz-Oaxaca). Estos dos experimentos consistieron en arreglos lineales de estaciones sísmicas de banda ancha perpendiculares a la trinchera con el fin de obtener extensos registros de formas de onda para caracterizar la subducción de la placa de Cocos. Mientras MASE se enfocó en el segmento plano de la zona de subducción en el estado de Guerrero, México, VEOX se enfocó en determinar la geometría de la placa inclinada en el Istmo de Tehuantepec. En este trabajo presentamos la red MIST que está siendo desplegada en el occidente de Michoacán, un área donde convergen procesos sísmicos y volcánicos. Estas tres redes tienen como objetivo conjunto caracterizar la geometría y reología de la placa en subducción, así como analizar los enjambres sísmicos asociados a los procesos magmáticos que originan el campo volcánico monogenético Michoacán-Guanajuato. En este trabajo se muestran los resultados de aplicar la técnica del cociente espectral entre las componentes horizontales y la vertical (HVSr) y su comportamiento direccional en bajas frecuencias para determinar si es posible inferir rumbo y buzamiento de la placa, además de la detección y análisis de tremores a lo largo del transecto, con énfasis en su relación con procesos volcanotectónicos. Por último, se delinean las perspectivas de expansión de la red y los análisis futuros orientados a integrar la dinámica tectónica de la región.

Palabras clave: Michoacán, México, H/V direccional, redes temporales, tremor, vulcanismo monogenético, subducción.

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1. Introduction

The state of Michoacán in Mexico is well known for being the birthplace of the Parícutín volcano one of the youngest monogenetic volcanoes of the world, the Parícutín volcano (Lühr and Simkin, 1993). Its coastal Pacific area is widely recognized as the epicenter of the $M_s=8.1$, 19 September 1985, Michoacan earthquake, the most catastrophic and deadliest seismic event in Mexico, until now (Beck and Hall, 1986; Esteva 1988; Campillo *et al.*, 1989). Despite holding large hazards (Sawires *et al.*, 2021 and 2023), fewer studies and seismotectonic experiments have been carried out here compared to other Mexican states along the Middle America Trench (MAT). In the past, large instrumentation extensive arrays were deployed in other states, such as Jalisco, Colima, Michoacán, Guerrero and Oaxaca. Between 2006 and 2007, with the objective of studying the Rivera plate subduction, the transition to the Cocos plate subduction, and the Jalisco Block, an array of 50 seismic stations was deployed in Jalisco, Colima, and western Michoacán, called the Mapping the Rivera Subduction Zone (MARS) array. To the Southeast, perhaps the most widely studied area in Mexico is the state of Guerrero which has been instrumented geodetically and seismically, both onshore and offshore. Between 2005 and 2007, a large seismic array consisting of 100 seismic stations provided a high-resolution seismic dataset. The so-called Middle America Subduction Experiment (MASE array) was deployed in a linear array perpendicular to the coast (Husker *et al.*, 2008; Pérez-Campos *et al.*, 2008), see Figure 1. The main goal of this experiment was to collect a large waveform dataset to finely image the geometry, rheology and tectonic history of the subducting Cocos plate. Additionally, a novel amphibious network deployed offshore the Guerrero coast (Cruz-Atienza *et al.*, 2018a) provided the first seismic and geodetic observations of the near-trench seismicity and deformation (Chen *et al.*, 2025; Cruz-Atienza *et al.*, 2021, 2025; Plata-Martínez *et al.*, 2021). A complementary experiment was deployed further east in the states of Oaxaca and Veracruz, called VEOX (Melgar *et al.*, 2011). This experiment examined the changes in geometry that produce drastic changes in the distribution of seismicity and the formation of volcanic field area ~385 km away from the trench (Espíndola *et al.*, 2016; Nelson *et al.*, 1995). Through these experiments, new insights about the past tectonic configuration of central and southeastern Mexico and their relationship with modern structures were established. However, there are still several issues that need to be understood such as the precise location and extension of seismic anomalies related to fluids and partial melt, and the accurate location of the slab interface and its mineralogy.

In this study, we present a new deployment of seismic station in the state of Michoacán, aimed at generating a dataset to map

the geometry of the subducting slab at the western flank of the flat segment of the Cocos plate. The western flank is geotectonically relevant due to the presence of the Michoacán-Guanajuato Volcanic Field (MGVF) that produces episodic seismic swarms, active volcanism and large subduction earthquakes. The Michoacan Seismic and Tremor (MIST) network was strategically set up to complete a long-term effort to investigate the subducting slab further.

In this contribution, we present the MIST network as a new observational framework to investigate a key unresolved problem in western Mexico: the link between slab geometry, crustal magmatic transfer, tremor-like seismic signals, and monogenetic volcanism in Michoacán. Specifically, we explore whether the first observations from the MIST array are consistent with: (1) along-strike changes in the geometry of the Cocos slab, (2) tremor-like signals associated either with slow slip or magma migration, and (3) the spatial relationship between active seismic swarms and the most hazardous sector of the Michoacán-Guanajuato Volcanic Field. This paper is therefore conceived as a preliminary-results and network report that establishes the scientific basis for future high-resolution studies.

2. Data and Instrumentation

In the first stage, ten seismic stations were deployed in a linear array roughly parallel to the MASE and VEOX networks, and nearly perpendicular to the strike of the MAT (see Figure 1). Furthermore, the layout was mainly established so that MIST array would complement the set of previous observations to image the western flank of the flat subduction interface. At the beginning of the project, three strong motion sensors and seven broadband stations were deployed starting in June 2023. Data availability and data gaps for each individual station are shown in [Figure S1](#). The array layout was chosen to cover an area that had not been previously instrumented by other experiments in the region such as the Mapping Rivera Subduction (MARS) temporary array and the Servicio Sismológico Nacional (SSN, National Seismological Service) permanent network. The total length of the MIST network extends approximately 270km, from the coastal city of Playa Azul to the town of San Cristobal de Ayala at the Michoacan-Guanajuato border, with an average interstation distance of ca. 30 km. The SSN IGIG station is located along the strike of the array about 50 km north of the northernmost sensor of MIST, this station was used as part of the network to extend the array's length to 320 km. The first three stations (AZUL, TAHU, PARE) included novel low-noise Silicon Audio optical accelerometers with data cube digitizers. The rest of the array were equipped with five broadband sensors Reftek

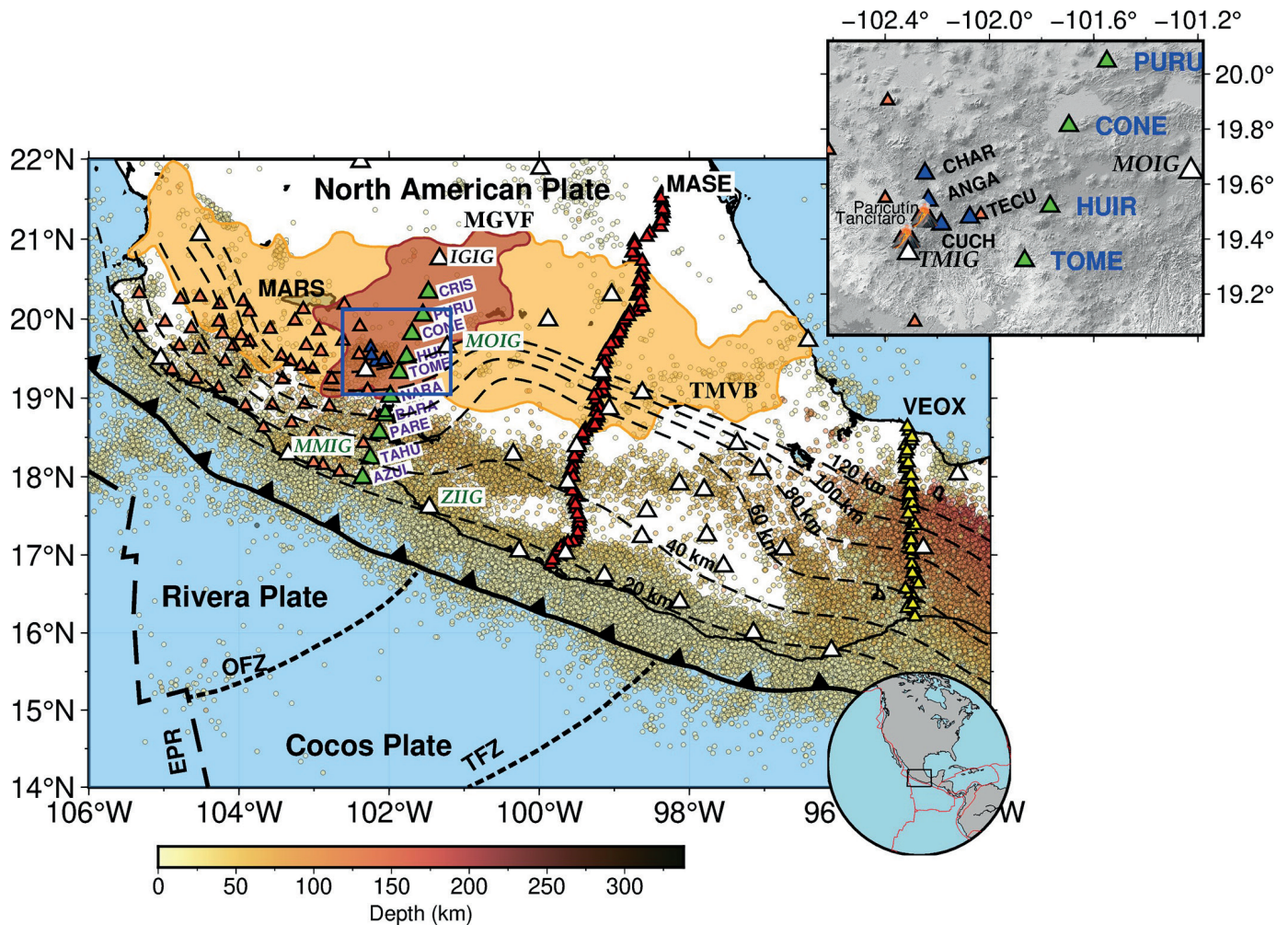


Figure 1. Map of stations of the MIST network and complementary arrays both temporary and permanent. The location of the MIST array is shown as green triangles and annotated with the four-character station codes. Circles color-coded by depth show the seismicity from 2000 to 2025 reported by the SSN (SSN 2025a). Orange triangles indicate the MARS (2006/01-2007/06) array. Location of the MASE array (2005/01-2007/05) and VEOX (2007/07-2009/03) arrays is shown with red and yellow triangles, respectively. Blue triangles indicate the location of Paricutín-Tancitaro Volcanic Monitoring Network (PT-VMN) (see upper inset). Broadband stations of the SSN (SSN, 2025b) are shown in white triangles; labeled stations correspond to those surrounding the MIST array. Orange and brown shaded area indicates the location of the Trans Mexican Volcanic Belt (TMVB) and the Michoacán-Guanajuato volcanic field, respectively (Sosa-Ceballos *et al.*, 2021). Blue square indicates the region in the top right inset. Rivera and Cocos plates boundaries are shown offshore. OFZ – Orozco Fracture Zone, TFZ – Tehuantepec Fracture Zone and EPR – East Pacific Rise. Dashed lines show the iso-contours of the subducting slab (Hayes *et al.*, 2018). Lower inset shows the global location.

130 sensors (BARA, TOME, HUIR, CONE), two Guralp 40T sensors (NARA, CRIS) and one Trillium Compact 120s sensor (PURU). All the originally installed sensors were later replaced by Guralp 3T sensors previously used in the MASE and VEOX experiments to collect broadband data for other studies such as receiver functions and seismic anisotropy. In all cases, we collected seismic records of 100 samples per second and decimated them according to the needs of each analysis.

A partner array of four stations was deployed west of the MIST array in collaboration among the United States Geo-

logical Survey (USGS), the Universidad Michoacana de San Nicolas de Hidalgo (UMSNH), and the Centro Nacional para la Prevención de Desastres (CENAPRED, National Center for Disaster Prevention). The Paricutín-Tancitaro Monitoring Network (PT-VMN) started operations in May 2024 and consists of four broadband 120 Trillium compact sensors located around the Tancitaro and Paricutín volcanoes. In addition to the seismic sensors, each station is equipped with geodetic and meteorological instrumentation. The distribution of this network is shown in the inset of Figure 1.

3. Subducted plate geometry in the state of Michoacán.

An ongoing question to solve in the Michoacán state is: What are the factors that determine the shape of the subducting slab and the elements that produce changes in its shape. Both global and regional studies suggest that the geometry of the subduction slab suddenly changes in the Michoacán region. Hayes *et al.* (2018) presented a global study of all active subducting zones around the world including Mexico. This study presents a general overview of the area that agrees well with regional studies such as those by Gaite *et al.* (2012) and Spica *et al.* (2016). In the eastern flank of the subduction zone, in the state of Oaxaca, Dougherty and Clayton (2014), Castellanos *et al.* (2018), Caló (2021) and Celis *et al.* (2024a, 2024b) imaged the fragmentation of the Cocos plate. In this case, the change in the Cocos plate geometry is most likely due to the subduction of the Tehuantepec ridge due to differential plate motion rates and the age of the slab. This observation is consistent with the shear-wave splitting tomography obtained by Celis *et al.* (2024a), which suggested a toroidal mantle flow around the dipping segment of the Cocos plate. In the case of Michoacán, the location of the Orozco fracture zone may be a factor to consider as a possible fragmentation mechanism in the west side of the flat segment, Dougherty *et al.* (2012) proposed that the fracture zone acts as barrier for the ultra-low velocity zone and therefore eases the bending of the slab. Furthermore, the younger and hotter slab suggests a more ductile regime that may explain the ductility of the subducting slab, in combination with a highly viscous mantle wedge as possible factors for the bending of the slab. The bending of the slab provides the thermochemical conditions needed for partial melting and the generation of monogenetic volcanism.

4. The Michoacán-Guanajuato Volcanic Field (MGVF)

The Michoacán state is characterized by two major geological provinces divided by the backarc basin called the Tepalcatepec depression (Garduño *et al.*, 1999). In the South, the Sierra Madre del Sur is a thrust belt composed of volcanic sedimentary rocks. In the central and northern sectors, it stands the largest monogenetic volcanic field in the world which holds around ~1000 volcanoes (Hasenaka and Carmichael, 1985). It consists of a few stratovolcanoes, cinder cones, maars, lava flows and domes (Hasenaka, 1994). The magmatism in the MGVF started during Early Pliocene, ~5 Ma (Guilbaud *et al.*, 2012) or ~6 Ma (Osorio-Ocampo *et al.*, 2018) with intense activity during the Holocene (Macías and Arce, 2019), including nowadays. This field includes the youngest continental volcanoes in Mexico, Jorullo and Parícutin (Guilbaud *et al.*, 2011; Luhr and Simkin 1993). One of the major landmarks in the Michoacán state is the

Parícutin volcano. This monogenetic cone emerged in a cornfield on 20 February 1942; its eruptive process lasted nine years until 1952 (Foshang and Gonzalez, 1956). Nonetheless, the largest volcanic structure in the state is the Tancítaro andesitic-dacitic stratovolcano. This volcanic structure rises ~2,000 meters above the surrounding terrain, reaching a maximum height of 3,840 meters above sea level (Ownby *et al.*, 2007). In the last 30 years, at least six seismic swarms have been reported around this area (1997, 1999, 2000, 2006, 2020 and 2021) (Caballero-Jiménez *et al.*, 2024; Legrand *et al.*, 2023; Mendoza-Rosas *et al.*, 2024; Pacheco *et al.*, 1999). Analysis of the seismic records from an ad hoc network installed weeks before the 2021 seismic swarm revealed a possible migration of the magma chamber towards the surface (Legrand *et al.*, 2023; Pertón *et al.*, 2024). Thus, a critical question to address is whether we can accurately forecast the emergence of a new volcano or the activation of the Tancítaro volcano. In the Parícutin–Tancítaro region, morpho-chronological analysis of cones using Average Erosion Index (AEI) and Voronoi tessellation indicates increasing rates of monogenetic volcanism over the last ~120 ka, with the highest probabilities located NE of Tancítaro, in strong agreement with the distribution of recent seismic swarms (Zarazúa *et al.*, 2024).

Figure 2 shows the reported seismicity by the SSN (SSN 2025a) from 2019 to July 2025. On these plots, three distinct seismic swarms are observed and shown in shaded areas in red, blue and purple (see Figure 2d). The first swarm of this seismic sequence occurred in the first quarter of 2020 being the most active in terms of the reported number of earthquakes per day, with a peak of up to 150 earthquakes per day above a completeness magnitude of M3.6. This swarm was exhaustively studied by Pertón *et al.* (2024). Using a local network these authors were able to identify nearly 100,000 earthquakes in two months with a successive reduction of the S-P time which suggested an upward migration of the seismicity. Earthquakes within this sequence have hypocentral depths mostly above 20km depth. A year later, a less energetic swarm reemerged at shallower depths, between 10 and 20 km (Legrand *et al.*, 2023). The 2021 swarm lasted about 3 months; though it was less energetic than the previous one (see Figure 2a, b, d). Finally, a third swarm, longer and more energetic than the 2020 and 2021 swarms, occurred between 2022 and 2023. This swarm was located northwest of the former sequences. During the activation of this swarm, a $M_w=4.6$ earthquake (2023/03/15) produced moderate damage in the nearby communities. Preliminary indicators of a new swarm began in late June 2025. Unlike previous crises, the swarm is being monitored by both the central segment of the MIST array (stations HUIR and TOME) and the Parícutín-Tancítaro Volcanic network. The analysis of this swarm is beyond the scope of this paper, but it will be addressed in subsequent works.

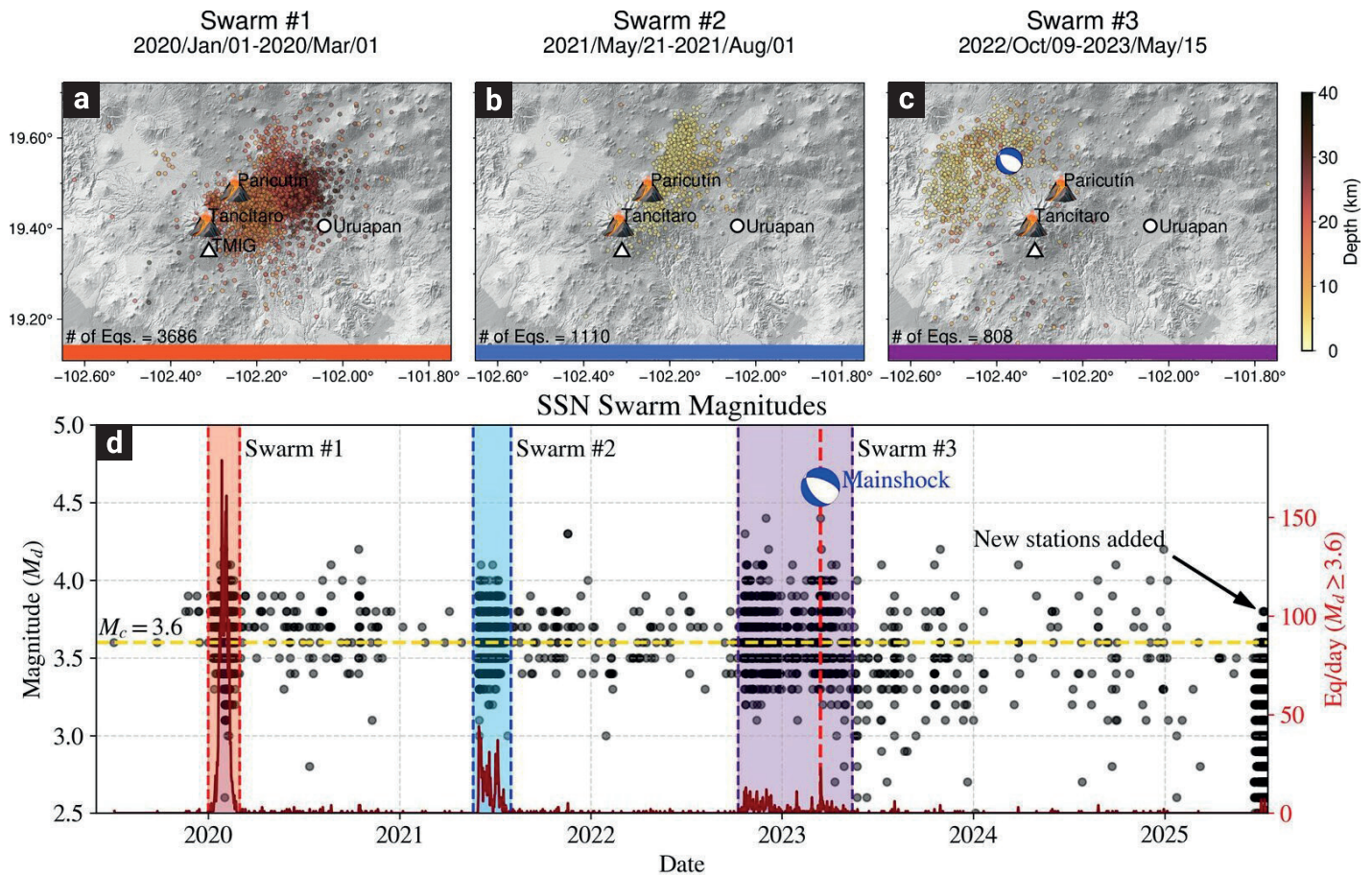


Figure 2. Reported seismicity from SSN from between January 2020 and July 2025/07. (a), (b) and (c) show map views of the earthquakes associated with three major seismic swarms. Earthquakes are color-coded by depth; the white triangle marks the location of the permanent SSN station TMIG, and the white circle indicates the location of the nearest major city, Uruapan. The colored bars at the bottom of each map (red, blue and purple) correspond to the time periods of the three swarms as shown in panel (d). Panel (d) shows the temporal variation and cumulative number of reported earthquakes. Circles represent reported earthquakes (SSN, 2025a), while the red line shows the number of earthquakes per day with magnitude $M_d \geq 3.6$ (horizontal yellow dashed line). The beachball shows the occurrence of the largest event ($M_w = 4.6$) during the 2022-2023 seismic swarm. The focal mechanism was computed using the method described in Franco *et al.* (2020).

5. Tectonic tremors and volcanic signals

Tectonic tremors in Mexico are primarily documented in the state of Guerrero, in the so-called sweet spot, approximately 215 km from the trench (Cruz-Atienza *et al.*, 2015 and 2018b; Husker *et al.*, 2012; Payero *et al.*, 2008; Maury *et al.*, 2016; Villafuerte and Cruz-Atienza, 2017) and in small clusters near the trench along the Guerrero Seismic Gap (Chen *et al.*, 2025; Plata-Martinez *et al.*, 2021). Numerous studies have proposed a strong connection between the aseismic slip and tremor activation (Ghosh *et al.*, 2009; Ito *et al.*, 2007; Wech *et al.*, 2007). In the east of Oaxaca both tremors and slow slip events (SSE) have been documented (Brudzinski *et al.*, 2010; Cruz-Atienza *et al.*, 2025b; Fasola *et al.*, 2016) with a relatively short recurrence time of 1-2 years. An obvious question is whether tremors may

occur in the state of Michoacán. If so, whether SSE events take place in this area or not. Brudzinski *et al.* (2016) reported the location of tremor and slow slip along an elongated band in the states of Colima and Jalisco. This study also reported some tremor activity, mainly in the western part of the Michoacán state. Until now, several studies targeting the detection of non-volcanic tremor agree that tremors are the result of aseismic slip of the plate interface between 40-60 kilometers (for example, Ide, 2012; Maury *et al.*, 2016; Husker *et al.*, 2019; Villafuerte and Cruz-Atienza, 2017).

We performed a spectral analysis for all stations of the MIST network during the first year of data. The analysis targeted the identification of narrow, high-energy bands in the 1–10 Hz range of the spectrogram. This feature is commonly associated with tectonic tremor and, more specifically, with the slow aseismic

slip of the plate (Beroza and Ide, 2011). Nonetheless, due to the complex tectonic setting and the existence of magma reservoirs, tremors may be associated with different phenomena. Figure 3 shows the map of stations and examples of possible tremor-like signals for two stations (NARA and BARA), as well as the time series for BARA station. The red rectangles in Figure 3a and 3b, indicate time windows where strips of energy in the 1-10 Hz band resemble to those observed in the state of Guerrero. Around this area the slab upper interface is expected to be at 60 to 80 km. Therefore, two possibilities may explain the existence of these signals on this specific region. First, this area roughly connects the tremors reported by Brudzinski *et al.* (2016) and Maury *et al.* (2018) in the state of Jalisco and the western segment of the state of Michoacán and the tremor sources in Guerrero (e.g. Cruz-Atienza *et al.*, 2015; Husker *et al.*, 2012; Payero *et al.*, 2008) in the flat segment of the slab. In this case, the activation of tremors would be an indicator of the existence of SSE at the interface, until now unreported. The second hypothesis suggests that these signals may be related

to magma migration associated with disseminated reservoirs in the crust. Under this assumption, tremors will not be connected to aseismic slip on the plate interface, but rather to the gradual upwelling of partially molten rocks and fluids. However, due to the station spacing and the volcanic nature of the region, large uncertainties remain regarding the origin of these signals. Consequently, it remains unclear whether the source of this activity is originated from magma migration or from aseismic slip processes. Based on present data, we favor the magma migration hypothesis for most of the tremor-like signals detected. The high dipping angle of the subducting slab in this region contrasts with the shallow angles in the Guerrero region where colder and pressurized fluid may favor aseismic slip at the interface as the main driving process for tremor radiation. If SSE does exist in Michoacán, the recurrence time interval is unknown, and it may be occurred at time intervals beyond the current observation window. If that is the case, the detection of SSE and tectonic tremors will be achieved once enough data is collected and GPS data becomes available.

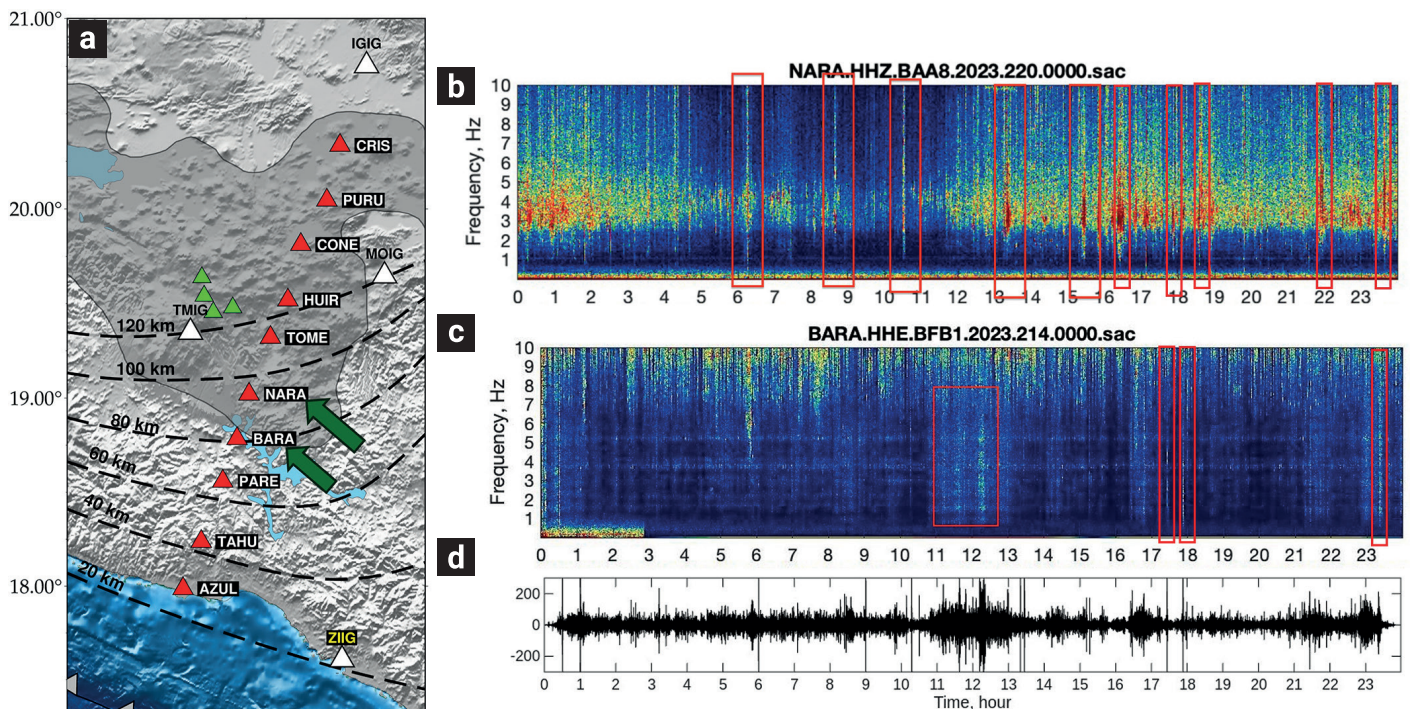


Figure 3. Location of the stations, spectral analysis and time series. (a) Map of stations including the MIST stations (red triangles), multi-parameter volcano-monitoring network (green triangles), and the SSN (white triangles) stations. The arrows indicate the two stations whose spectrograms are shown, NARA and BARA. (b) Spectrogram for the vertical component of station NARA (2 August 2023) and (c) spectrogram for east component of BARA station (7 August 2023). Red rectangles in (b) and (c) show possible tremors in the band of 1-10 Hz. (d) Original record at the BARA station bandpass filtered between 1-5 Hz.

6. Directional HVSR analysis

The horizontal-to-vertical spectral ratio technique (HVSR) was first proposed by Nagoshi and Igarashi (1971) and popularized by Nakamura (1989). This technique is widely used to assess the seismic site response based on the ratio between the horizontal and vertical components. A broad range of applications has been proposed over the years, such as the analysis of the structure at archeological sites (Piña-Flores *et al.*, 2021), volcanoes (Spica *et al.*, 2015), and sedimentary basin (Pertou *et al.* 2020) among others. In controlled experiments, it has been shown that the HVSR has strong azimuthal dependency. For example, Matsushima *et al.* (2017) observed large differences in the HVSR when using either EW or NS components. They observed that the azimuthal direction that provides the largest ratio between two rotated components correlates well with the location of a buried channel structure. Furthermore, Baena-Rivera *et al.* (2024) conducted a directional HVSR study along the lakebed zone near Mexico City, showing strong azimuthal dependence most likely related to lateral irregularities of the interfaces. These observations have been validated analytically (Sánchez-Sesma *et al.*, 2025) through numerical models that confirm the dependence of the angle that produces the largest difference in the estimates of the HVSR with the geometry of interfaces that induces this response. Here, we follow the methodology proposed by Baena-Rivera *et al.* (2024) to examine whether the HVSR at low frequencies below 0.03 Hz can support previous measurements of the strike and dip angle of the subducting slab. First, let us consider the standard formulation of the HVSR ratio,

$$\frac{H}{V}(f) = \sqrt{\frac{\langle |NS(f)|^2 \rangle + \langle |EW(f)|^2 \rangle}{\langle |V(f)|^2 \rangle}}, \quad (1)$$

where, H/V is HVSR, f is the frequency, and NS, EW, and V, correspond to the North-South, East-West and vertical components, respectively. $\langle |\cdot|^2 \rangle$ denotes the spectral densities average over time windows. Here, we follow the criterion proposed by Arai and Tokimatsu (2004) of computing the ratio of averages instead of the average of ratios, as the first way to compute it constitutes a solid theory for the HVSR based on the diffuse field assumption (Sánchez-Sesma *et al.*, 2011). This theory asserts that the ambient seismic noise (ASN) wavefield contains all kinds of elastic waves, understanding this as an elastic medium uniformly illuminated by equipartitioned energy.

The directional HVSR, which corresponds to the HVSRs computed with the horizontal movement components rotated between 0° and 180° , follows the next equations,

$$\frac{H'}{V}(\varphi, f) = \sqrt{\frac{\langle |H'(\varphi, f)|^2 \rangle}{\langle |V(f)|^2 \rangle}}, \quad (2)$$

where $H'(\varphi, f)$ is a horizontal rotated component an angle φ around the vertical axis from the EW direction, counterclockwise, using the 2D rotation transformation matrix,

$$\begin{bmatrix} H'_1(\varphi, f) \\ H'_2(\varphi, f) \end{bmatrix} = \begin{bmatrix} \cos \varphi & \sin \varphi \\ -\sin \varphi & \cos \varphi \end{bmatrix} \begin{bmatrix} NS(f) \\ EW(f) \end{bmatrix}. \quad (3)$$

Considering the two horizontal rotated components $H'_1(\varphi, f)$ and $H'_2(\varphi, f)$ are symmetric at 90° , that is, $H'_1(\varphi, f) = H'_2(\varphi \pm 90^\circ, f)$, only the use of one of the two horizontal rotated components, $H'(\varphi, f)$, is necessary. This means that they are complementary.

A measure of the amplitude difference between $H'(\varphi, f)$ and $H'(\varphi \pm 90^\circ, f)$ at a frequency of interest or in a range of frequencies gives us an indicator of the possible irregularities present in the interfaces of the soil in the site of ambient seismic noise recordings. This measure is the coefficient of the lateral irregularity, $\gamma(\varphi)$, computed using the following equation (Baena-Rivera *et al.*, 2024),

$$\gamma(\varphi) = \frac{1}{N} \sum_{f_1}^{f_2} \frac{|H'/V(\varphi, f) - H'/V(\varphi \pm 90^\circ, f)|}{\min(H'/V(\varphi, f), H'/V(\varphi \pm 90^\circ, f))}. \quad (4)$$

In a plane-layer case the directional HVSR does not depend on the rotation angle φ , then $\gamma(\varphi) \approx 0$ for every $\varphi \in [-90, 90]$, this indicates that the layers underneath the station at the depth expected for this frequency range are overall horizontal and planar. Conversely, in the case of the presence of lateral irregularities in the interfaces, i.e. non-planar layers or dipping, the directional HVSR depends on the rotation angle φ , then $\gamma(\varphi) \gg 0$. In summary, $\gamma(\varphi)$ is a measure of the lateral irregularity of an interface. This has been clearly tested for near-surface man-made structures in Onahana, Fukushima, Japan by Matsushima *et al.* (2017). The hypothesis has also been tested near Mexico City in the remains of a lakebed (Baena-Rivera *et al.*, 2024).

In this study, we assume that low enough frequencies (< 0.1 Hz) may provide insight into the deeper structure. Hence, we compute the directional HVSR for all the stations of the MIST array and four of the PT-VMN array using Eq. 2 and estimate $\gamma(\varphi)$ by applying Eq. 4. The directional HVSR was calculated following the diffuse field assumption (see Sánchez-Sesma *et al.*, 2011) using six days of ambient seismic noise velocity records, with 1000-s window lengths to ensure a frequency step of 0.001 Hz, and 10% overlapping. The STA/LTA criterion was

applied to avoid windows with spurious signals, transients, and monochromatic waves from anthropogenic sources. In addition, window normalization was applied before calculation using the unitary energy window criterion described in Baena-Rivera *et al.* (2024). Under the diffuse field assumption, the HVSR modeling is based on a spectral ratio of the windows averages for each principal direction, implying that the arithmetic mean of the energy represents the directional energy densities (DED). These DEDs are directly linked to Green's function and allow for a consistent inversion formulation. Therefore, in this study the uncertainty of the HVSR was performed by employing the bootstrapping method. This method takes the sample data and resamples it to create many simulated samples that have their own properties as mean and standard deviation. The method has an equal probability of randomly drawing each original data point for inclusion in the resampled datasets and can select a data point more than once for a resampled dataset. When applying the bootstrap method 200 HVSR were calculated by randomly selecting the first point of each time-average window of the original total signal. From the 200 HVSR, the HVSR mean ($\overline{H/V}$) and its standard deviation ($\sigma_{\overline{H/V}}$) was calculated.

Figure 4 shows the results for station CRIS (see location in Figure 1) as an example. The directional HVSR is shown in Figure 4a, where there are two important peaks, at low frequencies (below 0.03 Hz), susceptible to analysis. The first at 0.012 Hz with an amplitude of 178, and the second at 0.081 Hz with an amplitude of 10.4. Figure 4b shows the H/V total (black line) computed using Eq. 1, $H'/V(\varphi, f)$ (red line) and its counterpart $H'/V(\varphi + 90^\circ, f)$ (blue discontinued line) computed using Eq. 2. In this case, these last two curves are plotted for the rotation angles of 34° and 124° , respectively, where the maximum value for $\gamma(\varphi)$ is found for the frequency of interest. Also, the frequencies

of the peaks of interest and the value of $\gamma(\varphi)$ for those frequencies are indicated. Uncertainties in the estimation of the HVSR using the bootstrap method are shown in the Figure 4b.

At all stations, the same behavior of two main low-frequency peaks is observed: the first located between 0.005 and 0.02 Hz, and the second between 0.07 and 1 Hz.

From the φ at which $\gamma(\varphi)$ is maximum (i.e. a measure of the amplitude difference between $H'/V(\varphi, f)$ and $H'/V(\varphi + 90^\circ, f)$ at each station, and the value of $\gamma(\varphi)$, maps of the maximum lateral irregularity vectors of the directional HVSR like those in Figure 5 can be constructed. These types of maps can serve as support to detect the possible shape of interface irregularities.

Figure 5 presents two maps: the first (Figure 5a) shows the azimuthal behavior of the lateral irregularity vectors calculated between 0.005 and 0.02 Hz, where is located the first peak of the HVSR observed in Figure 4; and the second (Figure 5b) shows this behavior in the range 0.07 and 0.1 Hz, where the second peak of the HVSR observed in Figure 4 is located. The patterns of the two maps are similar, and few stations do not follow these patterns, such as CONE and PURU, which lack particularly good records. The common direction of the lateral irregularity vectors is the same.

Since the directional HVSR behavior is inherent to lateral variations and interface irregularities, it provides valuable information about structural complexity, thus the results can also be influenced by local heterogeneities, particularly in tectonically and volcanically active regions such as the study area. As shown in Figure 5, the slab strikes predicted by the HVSR technique (green arrows), yield strike values that are similar to those predicted by global models such as Slab2 (Hayes *et al.* 2018), mainly for those stations between the expected 40 km and 80 km isodepth lines (stations: TAHU, PARE, BARA). For the northern

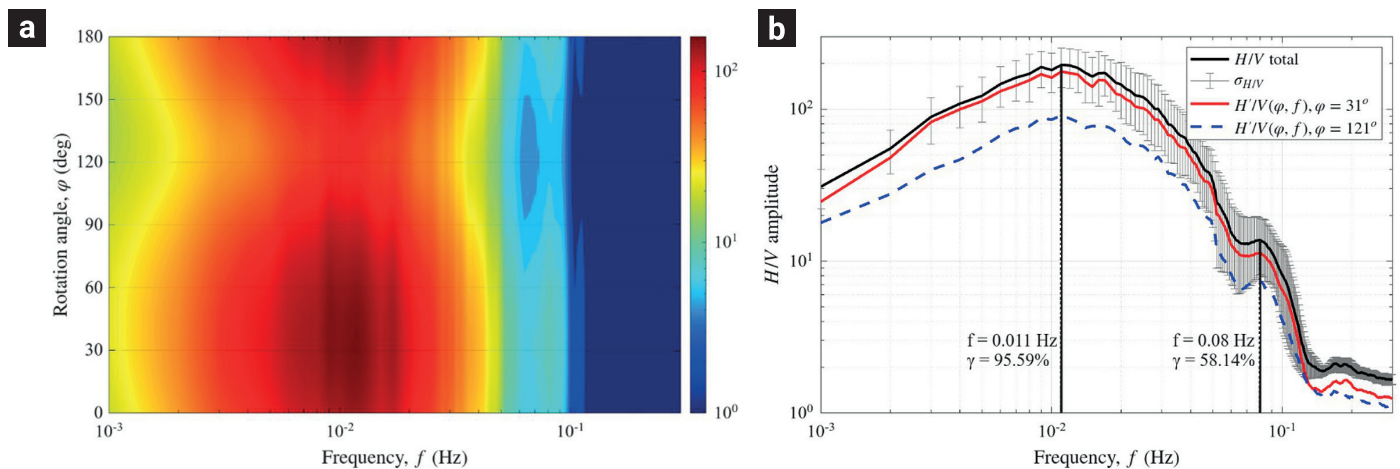


Figure 4. Directional HVSR for CRIS station of the MIST array. (a) Colormap of the directional HVSR. (b) H/V total and $H'/V(\varphi, f)$ for 34° and 124° rotation angles; the black thin vertical lines locate the frequencies of the two main peaks of the H/V total at low frequencies; error bars show the uncertainties within $\pm \sigma_{H/V}$; $\gamma \gg 0$ suggests a subsurface irregularity beneath the station.

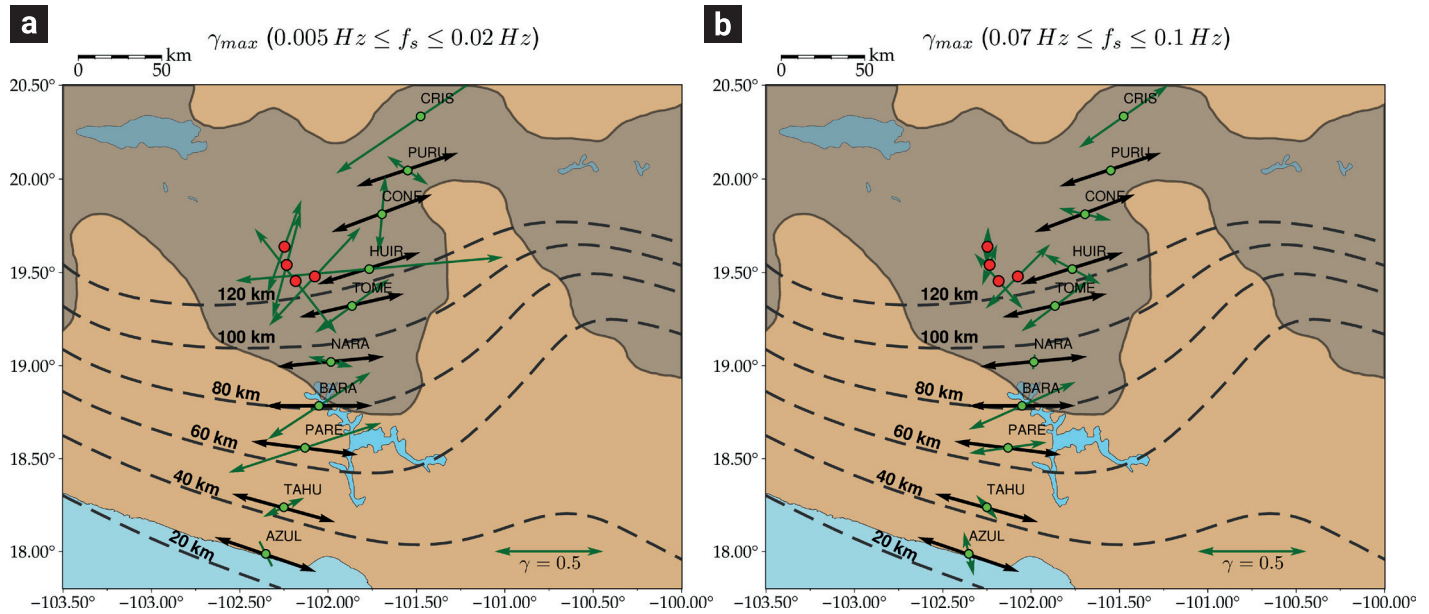


Figure 5. Azimuthal behavior of the maximum lateral irregularity vectors of the directional HVSR at MIST and PT-VMN stations, calculated at the peak frequencies of the HVSR between: (a) 0.005 and 0.02 Hz; and (b) 0.07 and 0.1 Hz. The sizes are their corresponding directionally dependent coefficient, $\gamma(\varphi)$. Dotted lines show the slab iso-depth lines from Slab 2 (Hayes *et al.* 2018), black arrows show the expected at each station. Green arrows show the predicted strike direction for the largest value γ in each frequency range. The length of the green arrows is scale by γ .

segment of the array (beyond NARA station), the predicted strikes become more randomly oriented. According to the Slab2 model, the slab in this area is either too deep (> 100 km) or it may not be even present at all. Therefore, the measurement may be most likely influenced by the crustal heterogeneity introduced by the presence of the TMVB. Interestingly, the predicted strikes for the PT-VMN are more coherent between stations and mainly oriented in the north-south direction. This is clearly a crustal effect due to the volcanic nature of the region while the estimates south of the TMVB are more likely due to sampling of average strike direction of the slab beneath the station.

7. Expansion plans and perspectives

Achieving high-resolution images of the subducting slab requires both a dense network with short inter-station distances (≤ 10 km) and relatively long recording periods (> 2 years). Given the current financial and instrumental resources, the MIST project will extend its recording window for the next two years (2025–2026). Five more stations will be deployed near the transition zone where the slab is expected to sink into the mantle. Figure 6 shows the proposed layout of stations, including the location of five 3-station temporary mini-arrays. This configuration is aimed at increasing the resolution of the studies. In the case of the station FELI, it is intentionally located off the array, to the

South, for a better location of the near-trench seismicity. The mini arrays will support seismic ambient noise-based studies and will fill the gaps between stations, except for the middle point between stations PARE and BARA due to extremely difficult conditions.

8. Conclusions

The MIST array represents an overdue project to understand better the variations in the geometry of the Cocos plate and its relationship with the formation of the Michoacán-Guanajuato Volcanic Field. The layout of this array was carefully chosen to mimic previous studies such as MASE and VEOX arrays. Although it is less dense in terms of the number of stations, the MIST array is recording enough broadband data needed to carry out multiple studies such as receiver functions, teleseismic and ambient noise tomography and volcano monitoring. In this paper, we presented a review of the past studies and the relevance of deploying the network in this area. One critical point to consider is the occurrence of periodic volcanic swarms. Some initial evidence of tremors and H/V directional properties along the network including a newly deployed 4-station broadband network in the Tancitaro volcano has been presented. First, we present the occurrence of earthquake swarms around the Tancitaro volcano. This phenomenon is currently a major concern among local communities and authorities, mainly because the

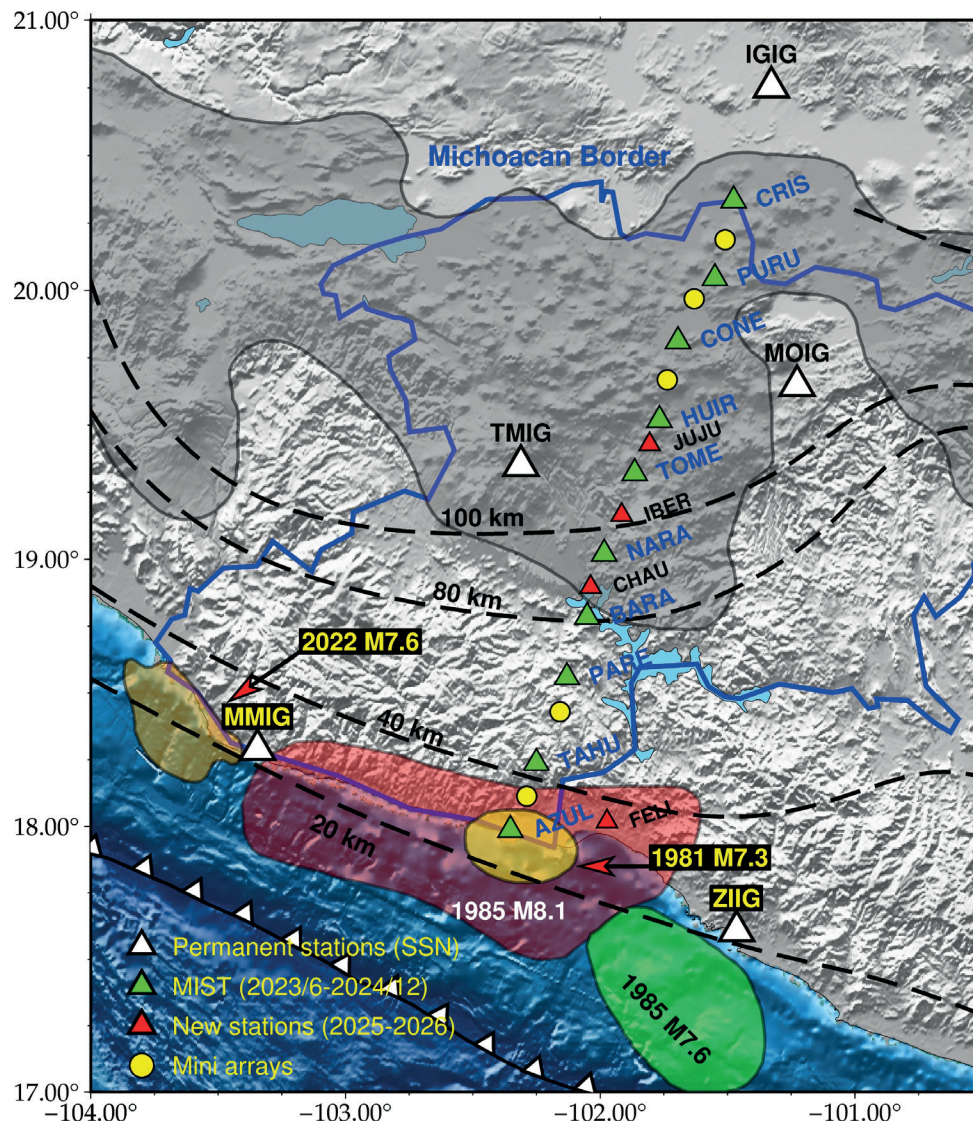


Figure 6. Future deployment of stations and temporary mini arrays. Color-shaded areas indicate the rupture areas of large earthquakes and the location of the TMVB (gray color shade), and the black dashed lines correspond to the iso-depths of the top of the slab according to Hayes *et al.* (2018). Green triangles show the location of the current MIST stations. Red triangle shows future deployment of five new stations. Yellow circles mark the location of five mini arrays of three stations each in between MIST stations for noise recording.

2023/03/15 $M_w=4.6$ earthquake that caused moderate damage. By including the central stations of the MIST array alongside the SSN and Parícutín-Tancítaro volcanic network, we will be able to determine the depth of the events with greater accuracy which is a fundamental parameter for assessing the volcanic hazard. Having a larger number of stations will provide a better constrain on the depth of the events and a better velocity model, which is a critical parameter to determine whether evacuations may be required. Additionally, we could better constrain the focal mechanism for relatively small earthquakes and monitor changes in V_p and V_s velocities. These factors are critical to properly characterize changes in the background seismicity and the properties of the medium. Furthermore, we found

tremor-like signals, although it was not possible to determine their origin due to the large interstation distance. Their possible existence suggests that SSE may also rupture within the state of Michoacán, forming a continuous band between the states of Guerrero and Jalisco where they had already been reported. The future deployment of additional stations will confirm or rule out whether this hypothesis holds or not.

Finally, we examined directional HVSR to evaluate whether this technique, at low frequencies, can determine the strike and dip of the subducting slab. Current estimates show promising results compared to global models and may provide an additional constraint to better model the geometry of the subducting slab. Moreover, the use of this technique also serves as a tool

to examine the complexity of the crustal structure. These initial results provide a baseline study for further instrumentation and investigation of the volcano-tectonic features in the state of Michoacán.

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Zona de cizalla Contreras, al sur de la Sierra de las Cruces: un ejemplo de la reactivación de fallas. Sector oriental del Cinturón Volcánico Transmexicano

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Abstract

The Sierra de las Cruces (SC) is a mountain range of volcanic origin located in the eastern sector of the Trans-Mexican Volcanic Belt. This range has an NNW-SSE trend and is composed of stratovolcanoes and lava domes of dacitic and andesitic composition of Pliocene-Pleistocene age. Although the SC is cut by a complex set of faults and has experienced recent seismic activity, aspects related to the kinematics, geometry, arrangement, and deformation history of these faults remain unknown. In this work, we analyze the NE-SW-trending faults, with emphasis on the Contreras Shear Zone. Our results reveal the existence of three families of faults in the SC, with NNW-SSE, ~E-W, and NE-SW trends. The ~E-W faults exhibit normal kinematics, while the NNW-SSE and NE-SW faults exhibit oblique kinematics with a dominant normal component. Normalized Correlation Sum (NCS) analyses indicate that faults in the CS have a cluster arrangement associated with the reactivation of previous structures. The Contreras shear zone has a NE-SW trend and has an exposed length of 35 km and a width of 10 km. It is also formed by a horst bounded by normal faults with a left-lateral component. The principal stress directions in this shear zone are 175/60, 078/04, and 345/29 for σ_1 , σ_2 , and σ_3 , respectively. Based on our results, we propose that the Contreras Shear Zone is the product of the reactivation of previous structures under three-dimensional deformation behavior.

Key words: Contreras shear zone, kinematics, fault reactivation, Sierra de las Cruces, Mexico basin.

Resumen

La Sierra de las Cruces (SC) es una cadena montañosa de origen volcánico ubicada en el sector oriental del Cinturón Volcánico Transmexicano. Esta sierra tiene una orientación NNW-SSE y está constituida por estratovolcanes y domos de lava de composición dacítica y andesítica del Plioceno-Pleistoceno. A pesar de que la SC es cortada por un complejo conjunto de fallas y que ha presentado actividad sísmica reciente, aún se desconocen aspectos relacionados con la cinemática, geometría, arreglo e historia de deformación de dichas fallas. En este trabajo se analizan las fallas con orientación NE-SW, poniendo énfasis en la zona de cizalla Contreras. Nuestros resultados revelan la existencia de tres familias de fallas en la SC, con orientaciones NNW-SSE, ~E-W y NE-SW. Las fallas ~E-W presentan cinemática normal, mientras que las fallas NNW-SSE y NE-SW muestran cinemática oblicua con una componente normal dominante. Los análisis de Suma Normalizada de Correlación (SNC) indican que las fallas en la SC tienen un arreglo agrupado asociado con la reactivación de estructuras previas. La zona de cizalla Contreras presenta un rumbo NE-SW y tiene una longitud expuesta de 35 km y un ancho de 10 km. Asimismo, está formada por un pilar tectónico limitado por fallas normales con una ligera componente lateral izquierda. Las direcciones de los esfuerzos principales en esta zona de cizalla tienen orientaciones de 175/60, 078/04 y 345/29 para σ_1 , σ_2 y σ_3 , respectivamente. De acuerdo con nuestros resultados, se propone que la zona de cizalla Contreras es producto de la reactivación de estructuras previas en condiciones de deformación tridimensional.

Palabras clave: Zona de cizalla Contreras, cinemática, reactivación de fallas, Sierra de las Cruces, cuenca de México.

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1. Introducción

La cuenca de México es una depresión continental limitada por cuatro serranías: al norte por la Sierra de Pachuca, al oriente por la Sierra Nevada, al sur por la Sierra de Chichinautzin y al poniente por la Sierra de Las Cruces (Figura 1. Arce *et al.*, 2019). Esta última sierra corresponde a una cadena volcánica que tiene una orientación preferencial NNW-SSE, la cual está constituida por domos y flujos de lava, estratovolcanes y complejos volcánicos formados durante el Plioceno-Pleistoceno y cuyas rocas son principalmente dacitas y andesitas (Figura 2; Arce *et al.*, 2019). Con base en las edades radiométricas (K-Ar y ^{40}Ar - ^{39}Ar) y variaciones en las polaridades paleomagnéticas, se ha sugerido que la actividad volcánica ha migrado de norte (3.7 Ma) a sur (0.9 Ma) (Aguirre-Díaz *et al.*, 2006a; Arce *et al.*, 2008; Mejía *et al.*, 2005; Mora-Alvarez *et al.*, 1991; Osete *et al.*, 2000). Las partes altas de la Sierra de Las Cruces están constituidas principalmente por lavas, mientras que sus laderas están conformadas por depósitos volcanoclásticos (lahares y flujos de escombros), piroclásticos (flujos piroclásticos y depósitos de caída) y fluviales. Todos estos depósitos se intercalan con sedimentos lacustres en las partes más bajas (Arce *et al.*, 2019).

La Sierra de Las Cruces está afectada por un conjunto de fallas geológicas con al menos tres orientaciones preferenciales: NNW-SSE, NE-SW y E-W (Figura 2.; García-Palomo *et al.*, 2008; Mooser, 1992). Aunque desde mediados del siglo XX se han descrito algunas de estas fallas, aún se desconocen la cinemática y la geometría de la mayoría de éstas. Asimismo, la relación entre los distintos sistemas de fallas es poco conocida, a pesar de que aparentemente han estado activas de manera simultánea desde el Pleistoceno tardío. Además, existe un lapso de más de 16 años en el que no se han realizado estudios estructurales detallados dentro de dicha sierra. Aunado a lo anterior, en los últimos años, la zona poniente de la Ciudad de México ha llamado la atención debido al fenómeno de los llamados “microsismos”. Esta actividad sísmica, caracterizada por eventos de magnitud menor a 3.3 y profundidades de entre 3 y 0.8 km (Servicio Sismológico Nacional [SSN], 2026), representa un peligro potencial para la población del poniente de la ciudad. Este fenómeno aceleró la necesidad de conocer con detalle la ubicación, cinemática e historia de actividad de las fallas geológicas localizadas en el sector occidental de la Ciudad de México, particularmente en la ladera oriental de la Sierra de Las Cruces.

La zona de cizalla Contreras, la cual se define por primera vez en este trabajo, constituye una estructura importante dentro del sector centro-sur de la Sierra de Las Cruces. Su prolongación hacia el noreste, dentro de la zona lacustre, intercepta la región donde se han registrado los microsismos. En este trabajo se analiza la cinemática, geometría e historia de actividad de esta zona de cizalla.

2. Marco Geológico

El Cinturón Volcánico Transmexicano (CVTM) es un arco volcánico activo desde el Mioceno (Ferrari *et al.*, 2012), originado por la subducción de las placas oceánicas de Cocos y Rivera por debajo de la placa Norteamericana (Figura 1). Tiene una longitud de más de 1000 km y un ancho variable de entre 90 y 250 km en promedio. A diferencia de otros arcos volcánicos, el CVTM no es paralelo a la trinchera debido a la subducción plana de la placa de Cocos, lo que le confiere una orientación preferencial E-W y provoca que, en su sector oriental, el arco esté muy alejado de la trinchera (Figura 1) (Ferrari *et al.*, 2012; Manea *et al.*, 2013).

Dentro del sector oriental del CVTM, se encuentra la cuenca de México (Figura 1). Esta cuenca está limitada por serranías de origen volcánico, las cuales se formaron en distintos periodos de actividad volcánica (Alba-Aldave *et al.*, 1996; Arce *et al.*, 2008, 2013, 2015; Bloomfield, 1975; Cadoux *et al.*, 2011; Fries, 1960; García-Palomo *et al.*, 2000; Jaimes-Viera *et al.*, 2018; Lenhardt *et al.*, 2010; Macías *et al.*, 2012; Mejía *et al.*, 2005; Morán-Zenteno *et al.*, 2004; Osete *et al.*, 2000; Siebe *et al.*, 2004a). Asimismo, la cuenca de México también es el resultado de la actividad de fallas geológicas, las cuales han estado activas desde el Oligoceno, como es el caso de la falla Mixhuca (Alaniz-Álvarez *et al.*, 2002). Otros grupos de fallas que moldean a la cuenca son el sistema de fallas Tenochtitlan NE-SW (De Cserna *et al.*, 1988; Vásquez-Serrano *et al.*, 2022), el sistema Apan NE-SW (García-Palomo *et al.*, 2018) y el sistema La Pera E-W (Delgado-Granados *et al.*, 1995).

2.1. Estratigrafía del sector centro-sur de la cuenca de México

Las rocas más antiguas dentro de la cuenca de México pertenecen al basamento sedimentario del Cretácico (Figura 1). Estas rocas son principalmente calcáreas y se agrupan dentro de las formaciones Morelos (rocas de plataforma), Cuautla (rocas de talud) y Mezcala (turbiditas de la cuenca de antepaís del Cretácico tardío). Las rocas calcáreas del basamento no afloran dentro de la cuenca; sin embargo, han sido cortadas por pozos a 1.5 y 2 km de profundidad (pozos Mixhuca y Tulyehualco respectivamente) y se exponen en el cañón de Lobos al oriente de Cuernavaca, Morelos (Figura 1. Arce *et al.*, 2019; Fries, 1960; Pérez-Cruz, 1988).

Sobreyaciendo discordantemente a las rocas calcáreas del Cretácico, se encuentra la formación Balsas del Eoceno (Figura 1), la cual se caracteriza por una intercalación de conglomerados, con clastos principalmente de calizas, depósitos volcanoclásticos, lacustres y fluviales (Fries, 1960). La formación Balsas no aflora dentro de la cuenca de México, pero ha sido descrita en pozos profundos (pozo Texcoco) y en el área de Cuernavaca

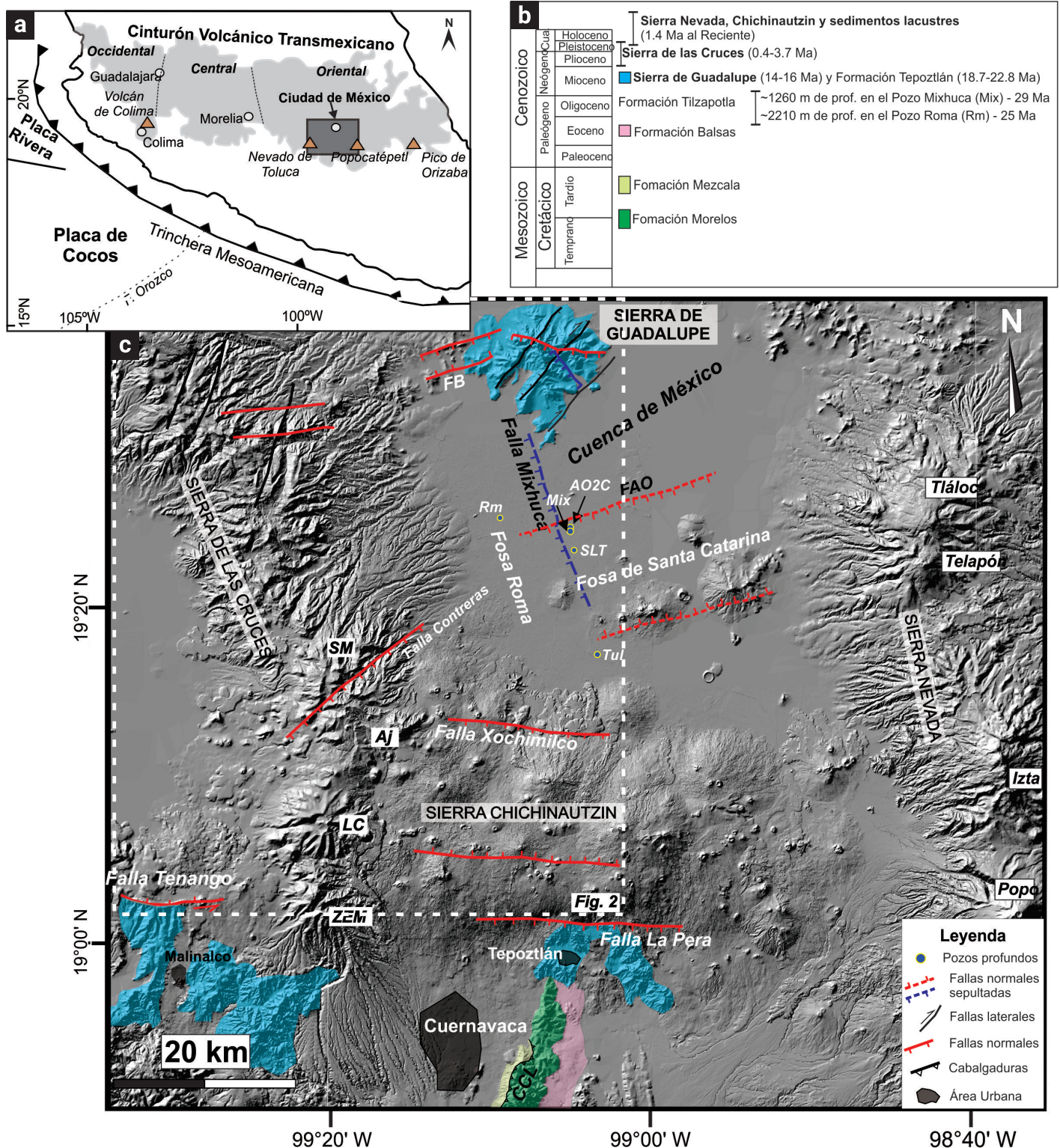


Figura 1. a) Mapa tectónico del Cinturón Volcánico Transmexicano, donde es posible apreciar a los principales estratovolcanes, así como la zona que delimita a la cuenca de México. b) Estratigrafía del sector sur de la cuenca de México. Nótese la posición estratigráfica de la Sierra de las Cruces. c) Mapa del sector centro-sur de la cuenca de México donde es posible apreciar las principales sierras que rodean a la cuenca, así como algunas fallas reportadas en trabajos previos (Pérez-Cruz, 1988; Norini *et al.*, 2006; Campos-Enríquez *et al.*, 2015; Arce *et al.*, 2019; Vásquez-Serrano *et al.*, 2022; Jaimes-Viera *et al.*, 2025), y la ubicación de algunos pozos profundos en la ciudad de México (Rm: Roma; AO2C: Agrícola Oriental 2C; Mix: Mixhuca; SLT: San Lorenzo Tezonco; Tul: Tulyehualco). SM: Volcán San Miguel; Aj: Volcán Ajusco; LC: Volcán La Corona; ZEM: Volcán Zempoala; Izta: Volcán Iztaccihuatl; Popo: Volcán Popocatepetl; FB: Fosa Barrientos; FAO: Falla Agrícola Oriental. CCL: Cabalgadura Cañon de Lobos. El recuadro de línea punteada es el área del mapa de la figura 2.

(Arce *et al.*, 2019; Fries, 1960). Cubriendo a la Formación Balsas se encuentra un grupo de rocas volcánicas de la formación Tilzapotla de edad Oligoceno (Figura 1). Dentro de la cuenca de México, estas rocas se han encontrado en los pozos profundos Texcoco, Mixhuca y Roma a profundidades de entre 950 y 2200 m (Arce *et al.*, 2019; Pérez-Cruz, 1988). Sobreyaciendo a las rocas volcánicas del Oligoceno se encuentra la formación Tepoztlán del Mioceno. Esta formación está constituida por una intercalación de lavas (andesitas y dacitas), depósitos piroclásticos, volcanosedimentarios y fluviales (Lenhardt *et al.*, 2010). Las rocas de esta formación se han ubicado en varios pozos profundos (Mixhuca, Texcoco, Copilco, Tulyehualco y San Lorenzo Tezonco), así como en la Sierra de Guadalupe, lo cual sugiere que está ampliamente distribuida dentro de la cuenca de México (Arce *et al.*, 2013, 2015, 2019, 2020; Pérez-Cruz, 1988; Vázquez-Sánchez y Jaimes-Palomera, 1989).

En la región centro-sur de la cuenca de México, sobre la formación Tepoztlán, existen depósitos principalmente volcánicos del Plioceno-Holoceno, relacionados con la actividad volcánica de las sierras: Nevada, Chichinautzin y Las Cruces (Figuras 1 y 2). La Sierra de las Cruces, con una orientación preferencial NNW-SSE y una longitud de ~80 km, está conformada por domos y flujos de lava, estratovolcanes y complejos volcánicos entre los que destacan La Catedral, La Bufo, Iturbide, Chimalpa, Salazar, San Miguel, Ajusco, La Corona y Zempoala (Figura 2). Estas estructuras volcánicas tienen un rango de edad de entre 3 y 0.4 Ma (Aguirre-Díaz *et al.*, 2006a; Arce *et al.*, 2008; Mejía *et al.*, 2005; Mora-Alvarez *et al.*, 1991; Osete *et al.*, 2000), y composiciones de andesita y dacita principalmente (Figura 2), aunque, Arce *et al.* (2019) reportan una edad de 37 ka para un depósito de caída asociado al estratovolcán San Miguel, ubicado en el centro-sur de la sierra. Por su parte, la Sierra Nevada, con una orientación N-S y una longitud de ~70 km, incluye a los estratovolcanes Tláloc, Telapón, Iztaccíhuatl y Popocatepetl (Figura 1). No existe un patrón en la variación de la edad de las estructuras volcánicas, pero es probable que el volcán Tláloc sea el más antiguo (1.8 Ma), mientras que el volcán más joven es el Popocatepetl, actualmente activo (Cadoux *et al.*, 2011; Espinasa-Pereña y Martin-Del Pozzo, 2006; Macías *et al.*, 2012; Rueda *et al.*, 2013; Siebe y Macías, 2004; Sosa-Ceballos *et al.*, 2012). La composición química de los productos volcánicos de la Sierra Nevada varía de andesitas a dacitas en los volcanes Iztaccíhuatl y Popocatepetl, y de andesitas a riolitas en los volcanes Tláloc y Telapón (Macías *et al.*, 2012; Martin Del Pozzo *et al.*, 2008; Nixon, 1989; Siebe *et al.*, 2017; Sosa-Ceballos *et al.*, 2012; Straub y Martin Del Pozzo, 2001). La Sierra de Chichinautzin, con una orientación E-W y una longitud de ~85 km, está conformada por al menos 220 volcanes monogenéticos con edades que van de 1.2 a 0.0016 Ma (Bloomfield, 1975). Con base

en su edad, la sierra de Santa Catarina, los volcanes Cerro de La Estrella y Peñón del Marqués, y el flujo de lava Papayo, también se agrupan dentro de la actividad de la Sierra de Chichinautzin (Arce *et al.*, 2019; Jaimes-Viera *et al.*, 2018; Macías *et al.*, 2012). Las composiciones químicas de los productos volcánicos asociados a esta sierra varían desde basaltos hasta dacitas (Arce *et al.*, 2013, 2015, 2024; Guilbaud *et al.*, 2009; Márquez *et al.*, 1999; Martin Del Pozzo, 1982; Meriggi *et al.*, 2008; Nieto-Torres *et al.*, 2023; Sandoval-García *et al.*, 2024; Siebe *et al.*, 2004b, 2005). Finalmente, intercalando y cubriendo los depósitos volcánicos y volcanoclasticos del Plioceno-Holoceno, se encuentra un paquete de sedimentos lacustres. Estos sedimentos tienen edades que van desde 1 Ma hasta el Holoceno y han sido descritos en algunos pozos profundos como en el San Lorenzo Tezonco y Tulyehualco (Arce *et al.*, 2013; Brown *et al.*, 2012; Caballero-Miranda y Ortega-Guerrero, 1998; Lozano-García y Ortega-Guerrero, 1998; Lozano-García *et al.*, 2017).

2.2. Estructuras en el sector centro-sur de la cuenca de México

La cuenca de México es una depresión, cuya geometría se debe a la presencia de fallas geológicas, las cuales controlan aspectos como su sedimentación, vulcanismo, sismicidad y subsidencia. Dentro de la cuenca se han identificado al menos tres sistemas de fallas principales que tienen orientaciones NNW-SSE, NE-SW y E-W. Aunque la relación temporal entre los distintos sistemas de fallas no es del todo clara, se ha propuesto que las fallas NNW-SSE y NE-SW registran actividad previa al Plioceno (Alaniz-Álvarez *et al.*, 2002; García-Palomo *et al.*, 2002, 2008, 2018; Vázquez-Serrano *et al.*, 2022). En contraste, las fallas E-W presentan un único registro de actividad dentro de la cuenca de México y podrían haberse originado bajo las condiciones tectónicas de extensión ~N-S en el sector oriental del CVTM (Figura 1. Alaniz-Álvarez *et al.*, 1998, 2002; Vázquez-Serrano *et al.*, 2022). No obstante, cabe mencionar que algunos estudios sugieren que las fallas E-W tuvieron actividad durante el Mioceno tardío (García-Palomo *et al.*, 2002).

Por otro lado, dentro de la cuenca de México destacan fallas kilométricas, como la falla Mixhuca con orientación NNW-SSE, la cual constituye el límite oriental de la fosa tectónica Roma (Figura 1. Pérez-Cruz, 1988). Otra falla kilométrica importante es la falla Agrícola Oriental, con orientación NE-SW (Figura 1), que representa el límite norte de la fosa Agrícola Oriental y forma parte del sistema de fallas Tenochtitlan (Arce *et al.*, 2019; De Cserna *et al.*, 1988; Vázquez-Serrano *et al.*, 2019). En cuanto a las fallas kilométricas con orientación E-W, se ha reportado la presencia de las fallas Xochimilco, Tenango y La Pera (Figura 1). Todas estas fallas se ubican dentro de la Sierra de Chichinautzin

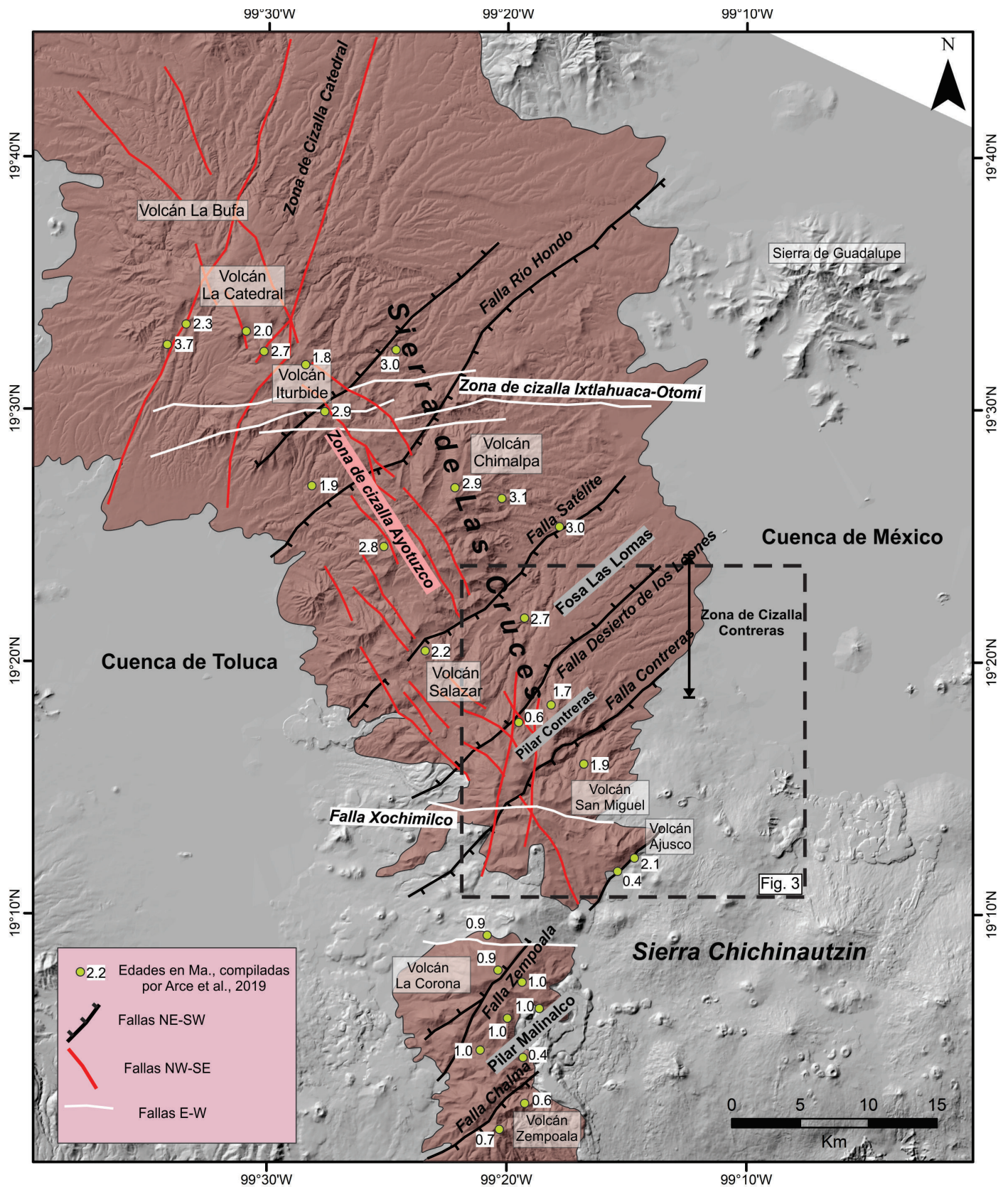


Figura 2. Mapa de estructuras de la Sierra de las Cruces, donde además se muestran los principales volcanes de la sierra (García-Palomo et al., 2008; Arce et al., 2019). En el mapa se incluyen las edades en millones de años (Ma) compiladas por Arce et al. (2019).

al sur de la cuenca de México y se agrupan dentro del sistema de fallas La Pera (Campos-Enríquez *et al.*, 2015; Delgado-Granados *et al.*, 1995; Lenhardt *et al.* 2010; Norini *et al.*, 2006). La actividad sísmica registrada en la cuenca de México sugiere que todos estos sistemas de fallas (NNW-SSE, NE-SW y E-W) son activos bajo el régimen extensional actual del CVTM (Bello-Segura *et al.*, 2025; Campos-Enríquez *et al.*, 2015; SSN, 2026).

2.3. Fallas geológicas en la Sierra de las Cruces

La Sierra de las Cruces está afectada por una serie de fallas geológicas pertenecientes a los tres sistemas reconocidos en la cuenca de México (García-Palomo *et al.*, 2008). Los pocos estudios existentes sobre esta temática son de carácter geológico-estructural, pero éstos no han logrado definir con precisión la geometría, cinemática, cantidad de desplazamiento ni la historia de actividad de muchas de estas fallas. En el trabajo de García-Palomo *et al.* (2008) se integró información de las fallas de toda la sierra y se identificaron algunas fallas kilométricas que dan una idea sobre la compleja deformación extensional dentro de la sierra (Figura 2). En general, estos autores describen tres sistemas de fallas con cinemática principalmente normal y orientaciones NNW-SSW, NE-SW y E-W. Entre las fallas NNW-SSE se encuentran la zona de falla Tula-Mixhuca y las fallas Ayotuzco; entre las fallas NE-SW destacan la zona de falla Catedral, la falla Río Hondo, la falla Satélite, la falla Contreras y la falla Chalma. Finalmente, las fallas E-W identificadas en la sierra son las fallas Ixtlahuaca-Otomí-Barrientos y las fallas Tenango-La Pera (Figura 2). Todos estos sistemas de fallas cortan la sucesión volcánica del Plioceno-Pleistoceno, lo cual sugiere que su actividad es principalmente holocénica.

3. Metodología

La metodología empleada en este estudio incluyó trabajo de campo en sitios estratégicos a lo largo de la traza de la zona de cizalla Contreras (Lomas de Tarango y Los Dinamos; ver figura 3). Durante el trabajo de campo se midió la orientación de las fracturas y fallas, se tomaron datos de estrías y se identificaron indicadores cinemáticos (p. ej., escalones, fibras de crecimiento, medias lunas y fracturas Riedel) con el objetivo de determinar la dirección de desplazamiento y el tipo de movimiento de las fallas. Asimismo, se realizó una descripción detallada de la zona de daño asociada a la falla Contreras (Figuras 3 y 4). Con los datos medidos en campo y tomando información previa de otros trabajos (Salinas-Garcés, 2014), se elaboró un mapa geológico-estructural de la parte centro-sur de la Sierra de las Cruces, que incluye los volcanes San Miguel y El Ajusco. Adicionalmente,

se construyó una sección geológica-estructural que muestra la geometría de la zona de cizalla Contreras.

El presente trabajo, también incluye la elaboración de un mapa de lineamientos morfológicos de la Sierra de las Cruces, tomando como referencia los trabajos de García-Palomo *et al.* (2008) y Vásquez-Serrano *et al.* (2022). Con este mapa se realizó un análisis de la distribución del tamaño de los lineamientos, así como de su arreglo, enfocándose principalmente en aquellos con orientación NE-SW. El arreglo de los lineamientos morfológicos y de las fracturas medidas en el campo se determinó mediante el código de Matlab CorrCount de Marrett *et al.* (2018), mientras que el análisis de estrías de falla, con el fin de obtener las direcciones de los esfuerzos principales y de extensión, se hizo con ayuda de los programas WinTensor (Delvaux y Sperner, 2003) y Faultkin (Allmendinger, 2025).

3.1. Arreglo de fracturas y distribución del tamaño

Para analizar el arreglo de las fracturas a escala kilométrica y mesoscópica, se midió su espaciamiento utilizando líneas de muestreo perpendiculares al rumbo de las estructuras (Figura 5). Para el caso de la escala kilométrica, con ayuda del mapa de lineamientos, se midió el espaciamiento entre lineamientos morfológicos a lo largo de toda la Sierra de las Cruces, mientras que, para la escala mesoscópica, se realizaron mediciones en campo dentro de la zona de daño de la falla Contreras (Figura 4), empleando cinta métrica y el comparador de Ortega *et al.* (2006). Los datos de espaciamiento fueron analizados en el código CorrCount de Marrett *et al.*, (2018), el cual compara el patrón de distribución espacial de las fracturas en una dimensión con patrones asociados con arreglos agrupados (distribución fractal), regularmente espaciados o aleatorios. Cada tipo de arreglo de fracturas puede estar relacionado con procesos de neoformación, reactivación y/o herencia de fracturas (Marrett *et al.*, 2018; Vásquez-Serrano *et al.*, 2024; Wang *et al.*, 2019).

Asimismo, se determinó la distribución del tamaño de las fracturas mediante la longitud (para los lineamientos morfológicos) y la apertura (a escala mesoscópica). Con estos datos se elaboraron diagramas de frecuencia-tamaño, a los cuales se les ajustaron tres funciones: ley de potencia, logarítmica y exponencial, utilizando una hoja de cálculo (Microsoft Excel®). De acuerdo con la literatura sobre análisis de fracturas, se ha sugerido que estas suelen mostrar un comportamiento fractal (Barton y La-Pointe 1995), por lo que su relación frecuencia-tamaño se ajusta a una función de ley de potencia, $y = x^n$, donde el exponente n es la dimensión fractal. En este trabajo, el ajuste en los diagramas frecuencia-tamaño se realizó considerando únicamente la parte central, donde los datos tienden a dibujar una recta. Los tramos que muestran tendencia a la horizontalidad (tamaños pequeños)

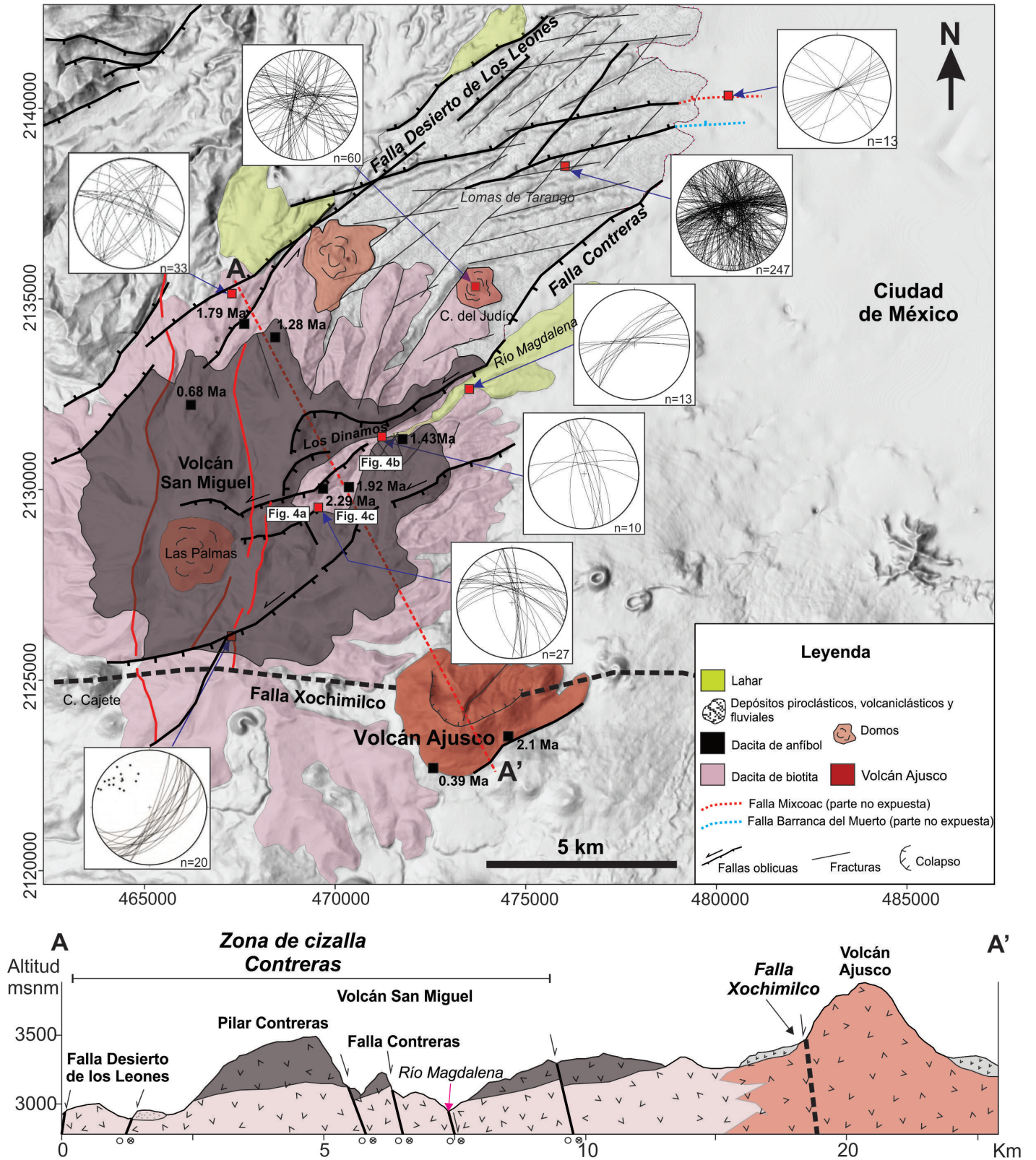


Figura 3. Mapa geológico-estructural y sección geológica de la zona de cizalla Contreras en el sector centro-sur de la Sierra de las Cruces. El mapa incluye datos de orientación de fracturas y fallas a lo largo de la zona de cizalla.

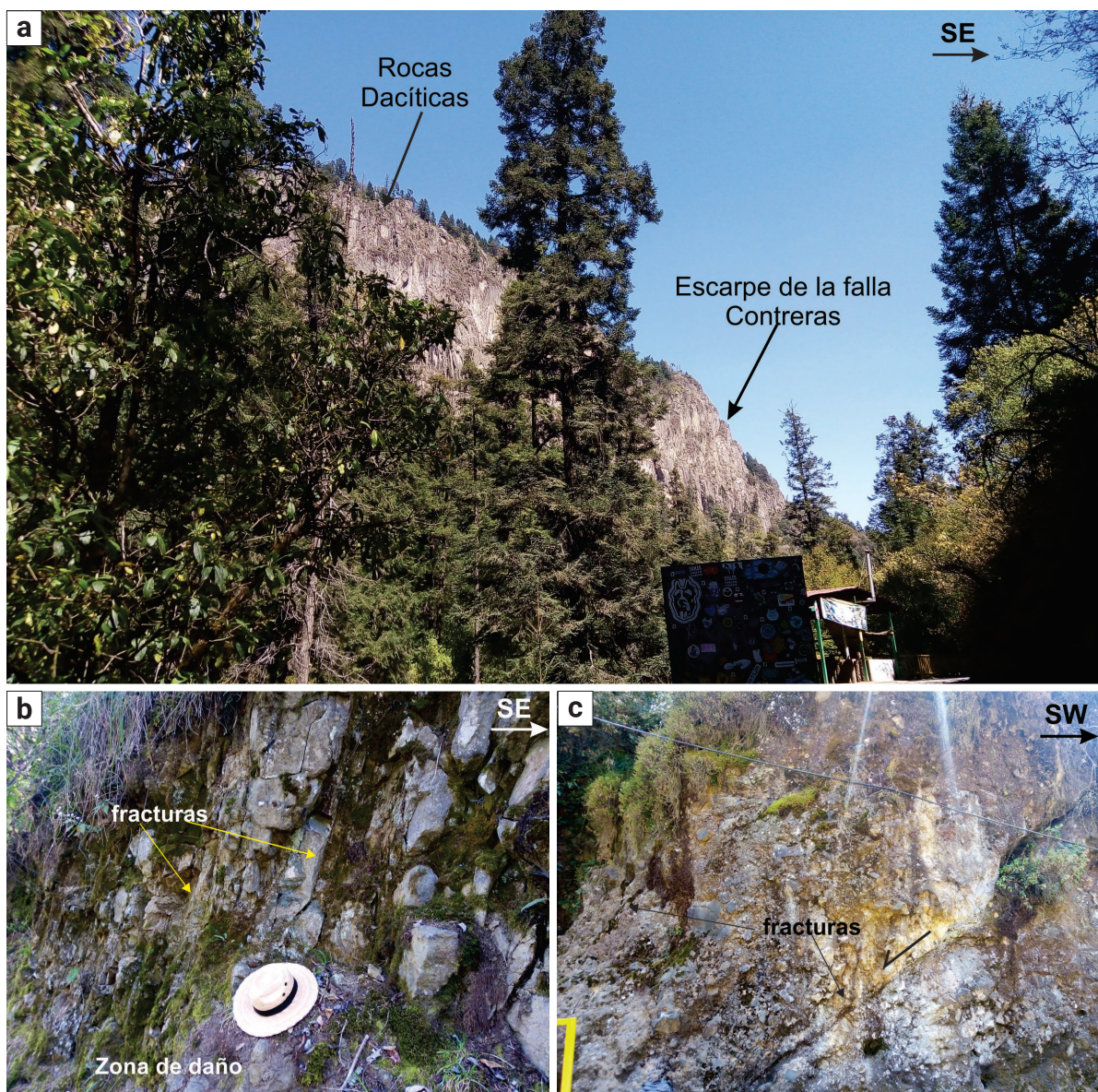


Figura 4. a) Vista panorámica del escarpe de la falla Contreras observada desde Los Dinamos, Ciudad de México. Nótese que la falla corta dacitas del Pleistoceno (ver figura 3). b) Zona de daño dominada por fracturas en el sitio Los Dinamos. c) Fracturas cortado a un depósito de lahar ubicado en el cañón del río Magdalena.

o a la verticalidad (tamaños grandes) no fueron considerados para calcular n , dado que se ha sugerido que reflejan efectos de censura en la medición de los datos, lo cual acarrea un error en la estimación de n (Barton y LaPointe, 1995).

4. Resultados

4.1. Definición de la zona de cizalla Contreras

La zona de cizalla Contreras se ubica en la región centro-sur de la Sierra de las Cruces, donde corta la parte central

del estratovolcán San Miguel (Figura 2). Esta zona de cizalla tiene un ancho de aproximadamente 15 km y está definida por un conjunto de discontinuidades kilométricas, segmentadas, anastomosadas y, en algunos casos, escalonadas, con rumbo preferencial NE-SW y algunas con rumbo ENE-WSW. Las discontinuidades con rumbo NE-SW presentan una cinemática oblicua con una componente menor lateral izquierda, y delimitan un pilar tectónico, denominado en este trabajo como “pilar Contreras”. El límite sur de este pilar está representado por la falla Contreras (Mooser, 1992), mientras que el límite norte está representado por la falla Desierto de los Leones (Figuras 2 y 3; García-Palomo *et al.*, 2008).

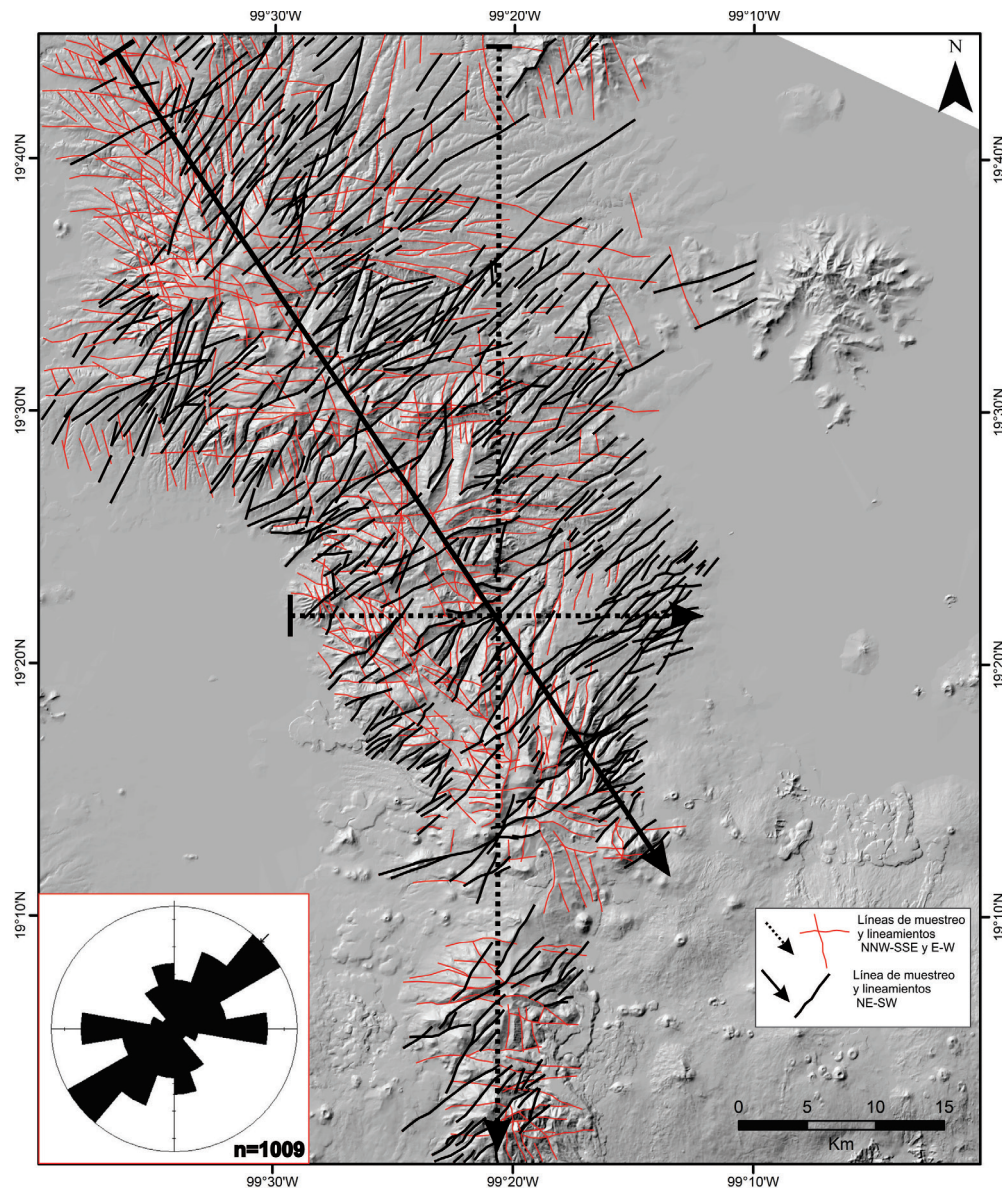


Figura 5. Mapa de lineamientos morfológicos de la Sierra de Las Cruces, donde se remarcan en color negro aquellos que tienen una orientación promedio NE-SW. Así mismo, se incluye la ubicación de tres líneas (Flechas) de muestreo perpendiculares al rumbo de las tres familias de lineamientos (NE-SW, E-W y NNW-SSE) mostradas en el diagrama de rosa en la parte inferior izquierda.

En la falla Contreras, algunos de sus segmentos son cóncavos hacia el SE y se trenzan a lo largo de una zona de alrededor de 2 km de ancho, mientras que en la falla Desierto de los Leones existen segmentos cóncavos hacia el NW, aunque los de mayor longitud suelen ser rectos (Figura 3). Dentro de la zona de cizalla Contreras, los segmentos de falla varían entre 0.5 y 10 km de longitud, y tienden a agruparse y a ser más numerosos en los límites del pilar Contreras, mientras que en la parte central del pilar son escasos o ausentes (Figura 3). Además, a lo largo de la traza de las fallas, en particular de la falla Contreras, se localizan algunas estructuras volcánicas de la Sierra de las Cruces y de la Sierra de Chichinautzin, como

los domos de lava Cerro del Judío y Las Palmas, así como el cono de escorias Cerro Cajete (Figura 3). En la región de Los Dinamos es posible observar el escarpe de la falla Contreras, donde corta a rocas dacíticas del volcán San Miguel (Figura 4). En esta región, el núcleo de la falla tiene un espesor de ~20 m, y una zona de daño caracterizada por una serie de fracturas de cizalla paralelas a la falla principal y fracturas extensionales con orientaciones ~E-W y ~N-S.

La zona de cizalla Contreras también incluye un grupo de fallas kilométricas que tiene un arreglo en mosaico, las cuales se ubican en el sector NE de dicha zona de cizalla (Figura 3). Las fallas tienen también una orientación NE-SW y ENE-WSW,

y afectan a depósitos volcánoclasticos, piroclásticos y aluviales en la zona de piedemonte de la Sierra de Las Cruces (Zona de Lomas) (Figura 3). En trabajos previos (Aguilar-Velázquez *et al.*, 2025), se han definido dos fallas con cinemática normal pertenecientes a este grupo, las cuales se han nombrado como: falla Mixcoac y falla Barranca del Muerto (Figura 3).

4.2. Lineamientos morfológicos en la Sierra de las Cruces

Dentro de la Sierra de las Cruces existe un conjunto de lineamientos morfológicos que tienen orientaciones preferenciales NNW-SSE, E-W y NE-SW. Esta última orientación es la más frecuente, mientras que la NNW-SSE es la menos común (Figura 5). Los lineamientos NE-SW están ampliamente distribuidos dentro de la sierra y tienden a formar agrupaciones incipientes, especialmente en zonas cercanas a las fallas kilométricas (Figuras 2 y 5). Por otro lado, los lineamientos NNW-SSE y E-W forman grupos mejor definidos, los cuales están localizados a lo largo de grandes zonas de cizalla (p. ej., zona de cizalla Ayotuzco o zona de cizalla Ixtlahuaca-Otomí; Figura 2).

Tomando en cuenta la longitud de los lineamientos morfológicos para cada familia (NE-SW, NNW-SSE y E-W) y representándolos en gráficos de frecuencia acumulada contra su longitud, se observa que las tres familias de lineamientos, así como todos los lineamientos (sin distinción de orientación), se ajustan a una función exponencial en la relación frecuencia-tamaño (Figura 6). No obstante, al eliminar las porciones horizontales y verticales de los gráficos de la figura 6, las cuales se han interpretado como resultado del sesgo en la definición de los lineamientos, los datos se ajustan mejor a una función de ley de potencia $y=x^n$. El exponente n obtenido para los lineamientos NE-SW es de -1.61, mientras que para los lineamientos E-W y NNW-SSE los valores son -1.45 y -1.55, respectivamente (Figura 6). Considerando el conjunto total de lineamientos, sin distinción de la orientación, se obtiene un valor de $n = -1.47$.

En el mapa de lineamientos de la figura 5 se trazaron tres líneas de muestreo perpendiculares al rumbo de las tres principales tendencias de lineamientos, donde se midieron datos de espaciamento y longitud. Con estos datos se realizaron gráficos que muestran la variación de la longitud de los lineamientos en función de la distancia (Figura 7). Los lineamientos NE-SW presentan una distribución relativamente homogénea, con una incipiente tendencia a la formación de grupos. Su longitud máxima es de 25 km y la mínima de 800 m. En contraste, los lineamientos E-W forman grupos mejor definidos, particularmente dos: uno al norte y otro al sur de la Sierra de las Cruces. Asimismo, su longitud máxima es de 30 km y la mínima de 2 km. Por su parte, los lineamientos NNW-SSE forman un patrón interesante en su distribución a lo largo de la línea de

muestreo E-W, donde muestran una tendencia a aumentar en su longitud hacia la parte central de la Sierra de las Cruces, donde se registra la máxima longitud (20 km). Las longitudes mínimas se encuentran en los extremos de la línea de muestreo, tanto al oriente como al poniente de la Sierra de las Cruces. A diferencia de las otras familias de lineamientos, en este caso no se observa una definición clara de grupos (Figura 7).

Aplicando el método de Suma Normalizada de Correlación (SNC) (Marrett *et al.*, 2018) a los datos de espaciamento entre lineamientos, se observa que las tres familias de lineamientos (NNW-SSE, E-W y NE-SW) presentan un arreglo característico de individuos agrupados (véase Figura 12e en Marrett *et al.*, 2018). Los grupos de lineamientos NE-SW tienen un ancho promedio de 8.7 km, mientras que en los grupos de lineamientos E-W y NNW-SSE los anchos son de 10 y 5 km, respectivamente. En todos los casos, la variación del SNC con respecto a la escala despliega un patrón de plataforma y cuenca (Figura 7). Estos patrones se definieron con mayor claridad en el código CorrCount al emplear un valor de 100 para “number of lags” y 50 para “windowing”.

4.3. Fracturas mesoscópicas en la zona de cizalla Contreras

Las fracturas mesoscópicas en la zona de cizalla Contreras son principalmente extensionales, de cizalla e híbridas. A escala mesoscópica, estas fracturas cortan a lavas dacíticas del estratovolcán San Miguel (Figura 4b), mientras que, en el piedemonte oriental de la sierra, dichas estructuras cortan a depósitos volcánoclasticos, piroclásticos y paleosuelos (Figura 3 y 4). Las fracturas presentan tres orientaciones preferenciales: NNW-SSE, E-W y NE-SW (Figura 3). En la región del piedemonte se observa que las tres familias de fracturas son principalmente extensionales e híbridas.

En el cañón de Los Dinamos, a lo largo de la traza de la falla Contreras (Figura 3), se identificó el conjunto de fracturas de Riedel (Riedel, 1929), las fracturas NNW-SSE son extensionales (T en el modelo de Riedel), mientras que las fracturas NE-SW y ESE-WNW son de cizalla. Las fracturas NE-SW son paralelas a la falla Contreras y representan a las fracturas tipo X, mientras que las fracturas ESE-WNW estarían asociadas con las discontinuidades tipo P. Dentro de la zona de daño de la falla Contreras se midió el espaciamento y la apertura (espesor en el caso de las fracturas de cizalla) de las fracturas extensionales (T) y de cizalla (X) en dos líneas de muestreo. Con estos datos se analizó la relación frecuencia-tamaño, la distribución de las fracturas a escala mesoscópica y se determinó su arreglo. Los gráficos de frecuencia acumulada contra el tamaño muestran que la variación del tamaño (espesor) de las fracturas de cizalla se ajusta mejor a una función logarítmica

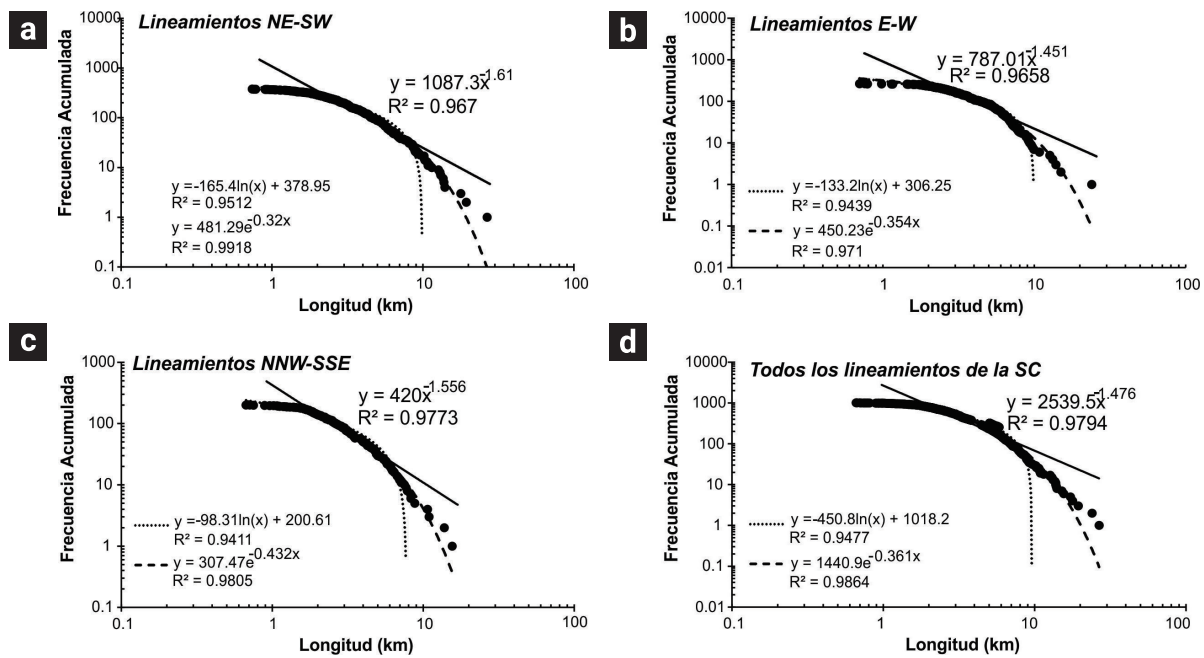


Figura 6. Gráficos de frecuencia acumulada contra la longitud para las tres familias de lineamientos (a, b, c) y todas en su conjunto (d). En cada gráfico se ajustan tres funciones: logarítmica, exponencial y de potencia. En el caso de la función de potencia, ésta se ajustó tomando en cuenta solo la parte central de los datos como se sugiere en Barton y LaPointe (1995).

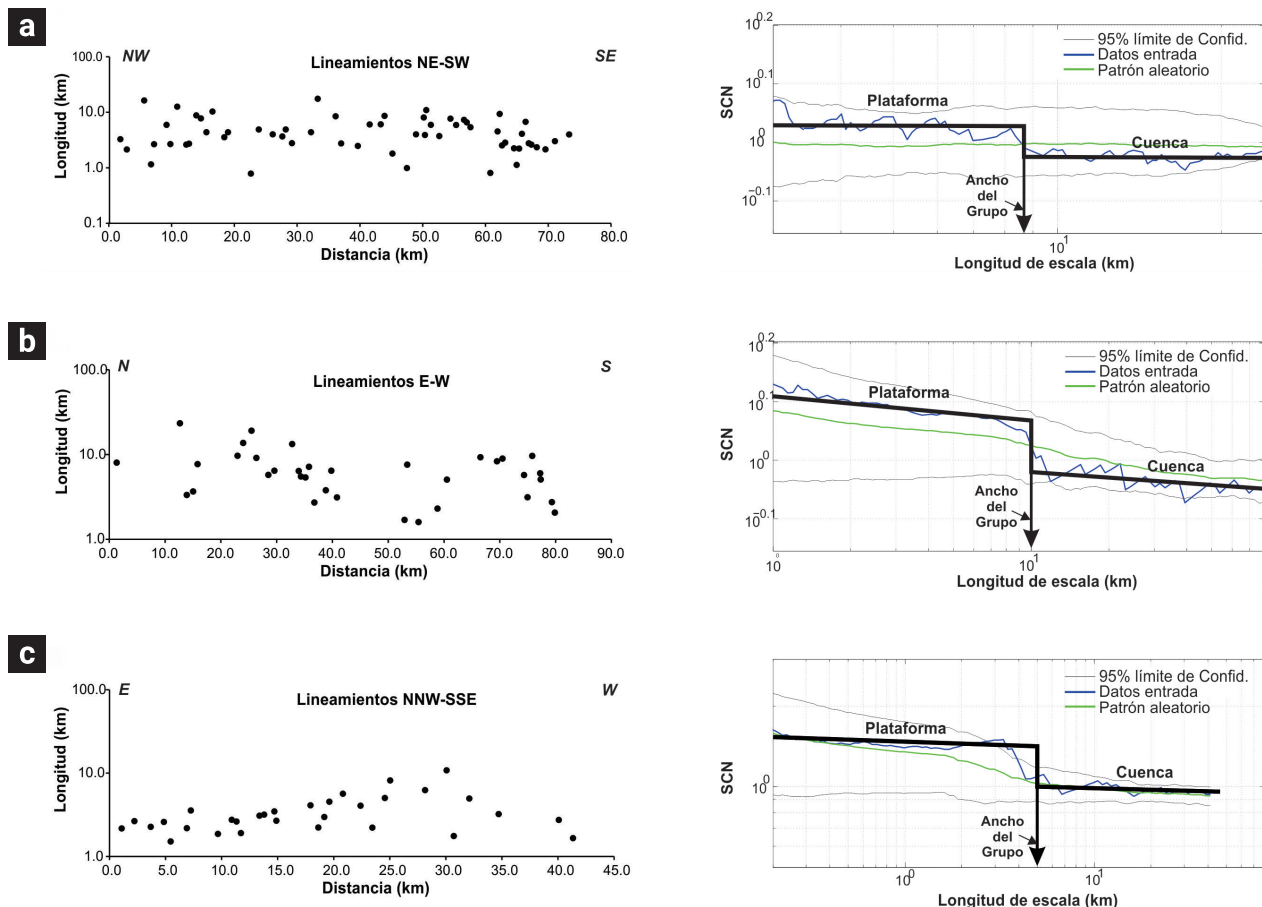


Figura 7. Gráficos de distribución espacial y diagramas de Suma de Correlación Normalizada (SNC) para los lineamientos NE-SW (a), E-W (b) y NNW-SSE (c). La distribución de lineamientos muestra que tienden a agruparse, lo cual es confirmado con el gráfico SNC, que sugiere un patrón de individuos agrupados (Marrett et al., 2018) para cada familia de lineamientos.

cuando se consideran todos los datos (Figura 8). No obstante, al eliminar el sesgo, la función que mejor describe la variación del tamaño es una ley de potencia, con un exponente n de -0.77 (Figura 8a). En el caso de las fracturas extensionales, los datos definen claramente dos pendientes en el gráfico frecuencia acumulada contra la apertura (Figura 9b), ambas ajustadas a una ley de potencia. Para las fracturas de mayor apertura (entre 0.3 y 0.8 cm), el exponente n tiene un valor de -2.37 , mientras que para las fracturas de menor apertura (entre 0.008 y 0.2) el exponente n es de -0.37 (Figura 8b).

La distribución de las fracturas de cizalla define ciertos grupos, los cuales incluyen a fracturas de diferente espesor (apertura) en un rango que va desde 0.01 a 4 cm (Figura 8c). Las fracturas extensionales también tienden a agruparse, aunque es notable

la formación de dos patrones: uno asociado con un rango menor en la apertura (0.1 a 0.8 cm) y otro relacionado con un rango más amplio (0.008-0.8 cm) (Figura 8d). Utilizando el método de SCN (Marrett *et al.*, 2018), es posible apreciar que las fracturas extensionales y de cizalla presentan arreglos distintos (Figuras 8e y f). En las fracturas de cizalla, el arreglo corresponde a grupos fractales (véase Figura 12f en Marrett *et al.*, 2018), en donde el ancho de los grupos es de 2 m. En las fracturas extensionales, en cambio, se observa un arreglo combinado: a escalas pequeñas (<60 cm) el patrón de individuos agrupados define mejor su arreglo, mientras que a escalas mayores (entre 60 cm y 2 m) predominan grupos fractales (Figura 8f). El ancho de los grupos en escalas pequeñas es de 55 cm, mientras que en escalas grandes es de 2 m (Figuras 8e y f).

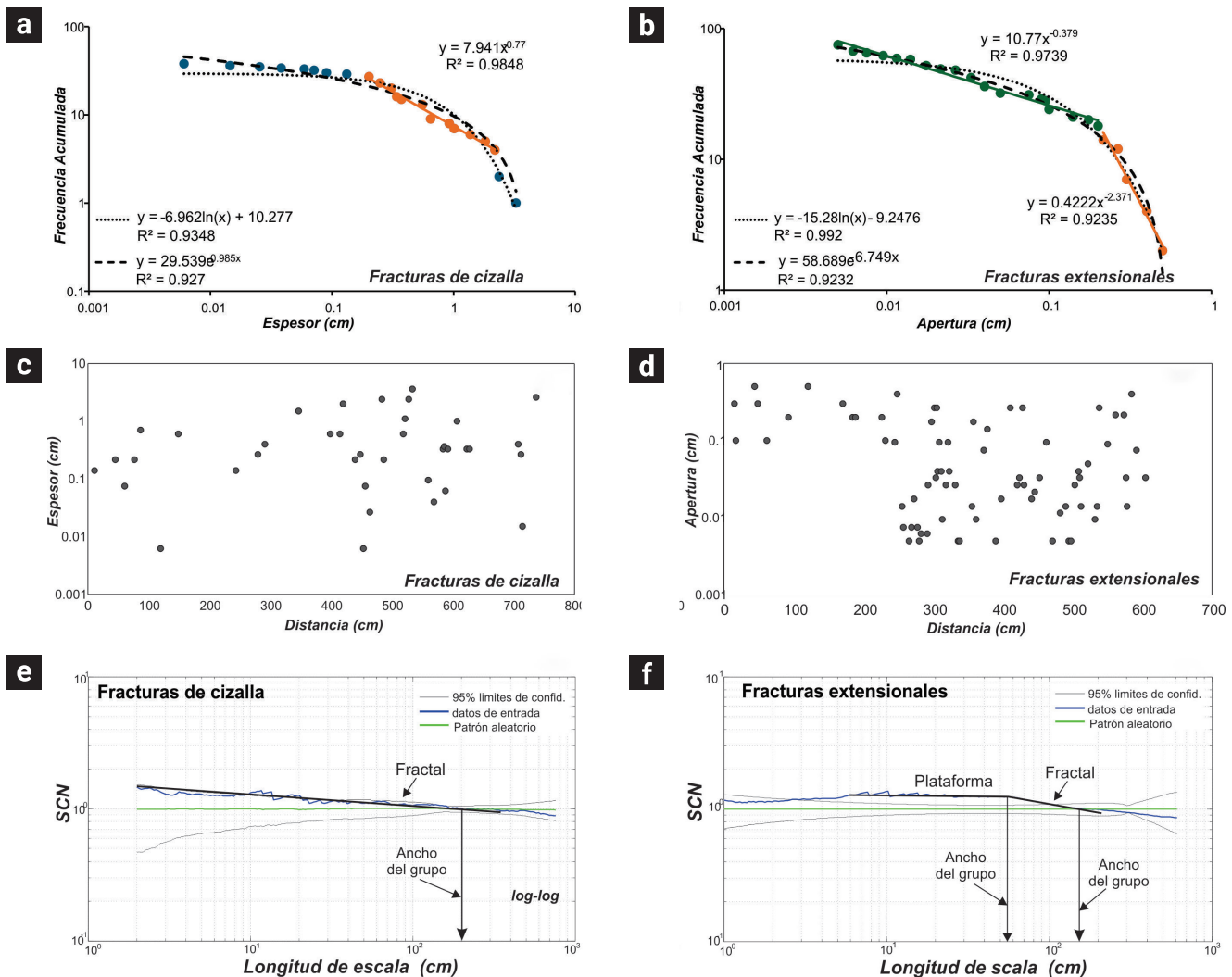


Figura 8. Gráficos de frecuencia acumulada contra el espesor y la apertura para fracturas de cizalla (a) y extensionales (b) dentro de la zona de daño de la falla Contreras. c, d) Distribución de fracturas a lo largo de las líneas de muestreo analizadas a escala mesoscópica. e, f) Diagramas de Suma de Correlación Normalizada (SCN) para las fracturas de cizalla y extensionales. Nótese que en el caso de las fracturas de cizalla se tiene un arreglo agrupado asociado con una distribución fractal, mientras que para las fracturas extensionales se forman dos arreglos, uno asociado con individuos agrupados (plataforma) y otro con una distribución fractal.

4.4. Análisis cinemático de la zona de cizalla Contreras

A lo largo de la zona de cizalla Contreras se midió la orientación, el vector de desplazamiento y el tipo de movimiento de varios planos de fallas mesoscópicas, con orientaciones NE-SW, NNW-SSE y E-W. Estos datos fueron analizados mediante el método de Bingham, utilizando el programa Faultkin (Allmendinger, 2025), con el objetivo de determinar la orientación de las extensiones principales (ϵ_1 , ϵ_2 , y ϵ_3). Se obtuvieron valores (rumbo/echado) de 342.2/19, 074/7.2, y 184.5/69.5 para ϵ_1 , ϵ_2 , y ϵ_3 , respectivamente. Estas orientaciones se asocian con una extensión NNW-SSW ligada con la zona de cizalla Contreras en el sector centro-sur de la Sierra de las Cruces. Asimismo, la pelota de playa obtenida a partir de los datos sugiere un meca-

nismo de falla de tipo normal con una ligera componente lateral izquierda, mientras que el diagrama de polos de los planos de fallas no describe sistemas conjugados (Figura 9a).

Por otro lado, con datos de orientación medidos en fracturas Riedel identificadas a lo largo de la falla Contreras (Figuras 3 y 9b), se determinó la forma del elipsoide de esfuerzos y la orientación de los esfuerzos principales, siguiendo la metodología de Angelier (1989) y utilizando el programa WinTensor. El proceso de inversión de datos se realizó mediante el método del *Right Dihedron*, obteniéndose una orientación (rumbo/echado) de 169/52, 264/04 y 357/38 para σ_1 , σ_2 , y σ_3 , respectivamente, mientras que R (relación de aspecto del elipsoide de esfuerzo) tuvo un valor de 0.41 (Figura 9b). Asimismo, el eje principal de acortamiento máximo presenta una orientación NW-SE en la

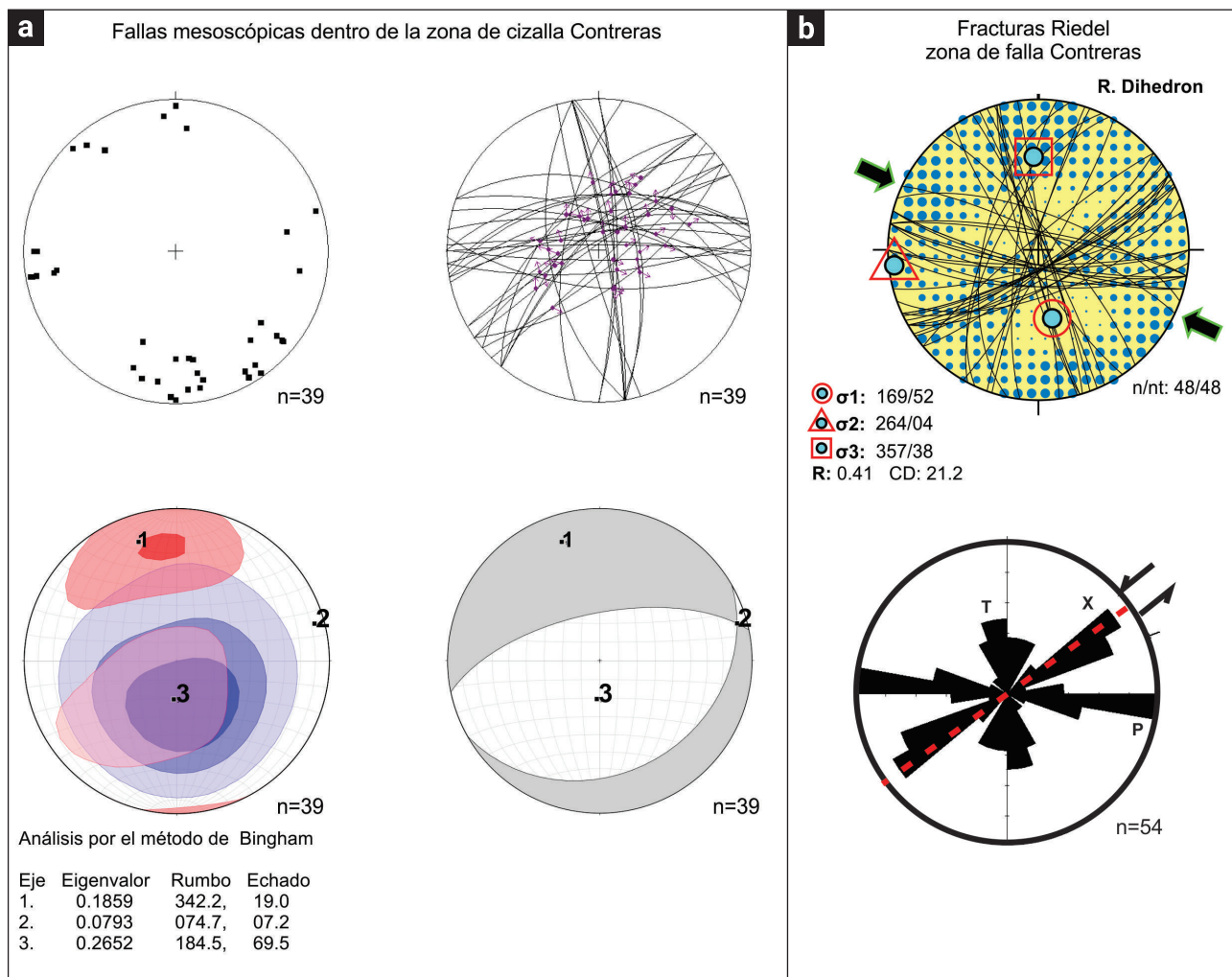


Figura 9. a) Proyección estereográfica de fallas mesoscópicas medidas dentro de la región centro-sur de la Sierra de las Cruces. Así como, diagrama de polos (para mostrar que no siguen un patrón romboédrico), pelota de playa (mecanismo promedio de falla) y direcciones de extensión (usado el método de Bingham) para las fallas analizadas. b) Dirección de los esfuerzos principales (σ_1 , σ_2 , y σ_3), proyección estereográfica y diagrama de rosa de facturas T y X, y P encontradas en la zona de daño de la falla Contreras. La dirección de los esfuerzos principales se obtuvo mediante el método *R. Dihedron* (Angelier y Mechler, 1977) en el programa WinTensor.

parte central de la falla Contreras (Figura 9b). Esta orientación coincide con las relaciones angulares entre las fracturas del conjunto Riedel, donde se observa que las fracturas extensionales (T) forman un ángulo mayor a 45° con respecto a las fracturas X, lo que sugiere un régimen local de transpresión en una falla oblicua con una componente lateral izquierda (Figuras 3 y 9b).

Ahora bien, si se toman en cuenta las fallas mesoscópicas dentro de la zona de cizalla Contreras y se aplica un proceso de inversión con el programa WinTensor, considerando el método de *Right Dihedron*, es posible observar en el gráfico de

Hoeppener que existen grupos de datos que tienen diferentes direcciones de extensión (Figura 10a). Por otro lado, la orientación de los esfuerzos principales obtenida muestra que σ_1 , σ_2 , y σ_3 tienen valores de 175/60, 078/04 y 345/29, respectivamente, y que el valor de R es de 0.33. Asociada con la orientación y la dirección de desplazamiento de las fallas mesoscópicas en la zona de cizalla Contreras, la extensión en el sector centro-sur de la Sierra de las Cruces tiene una orientación NW-SE, aunque también existe una extensión menos intensa en la dirección NE-SW (Figura 10b).

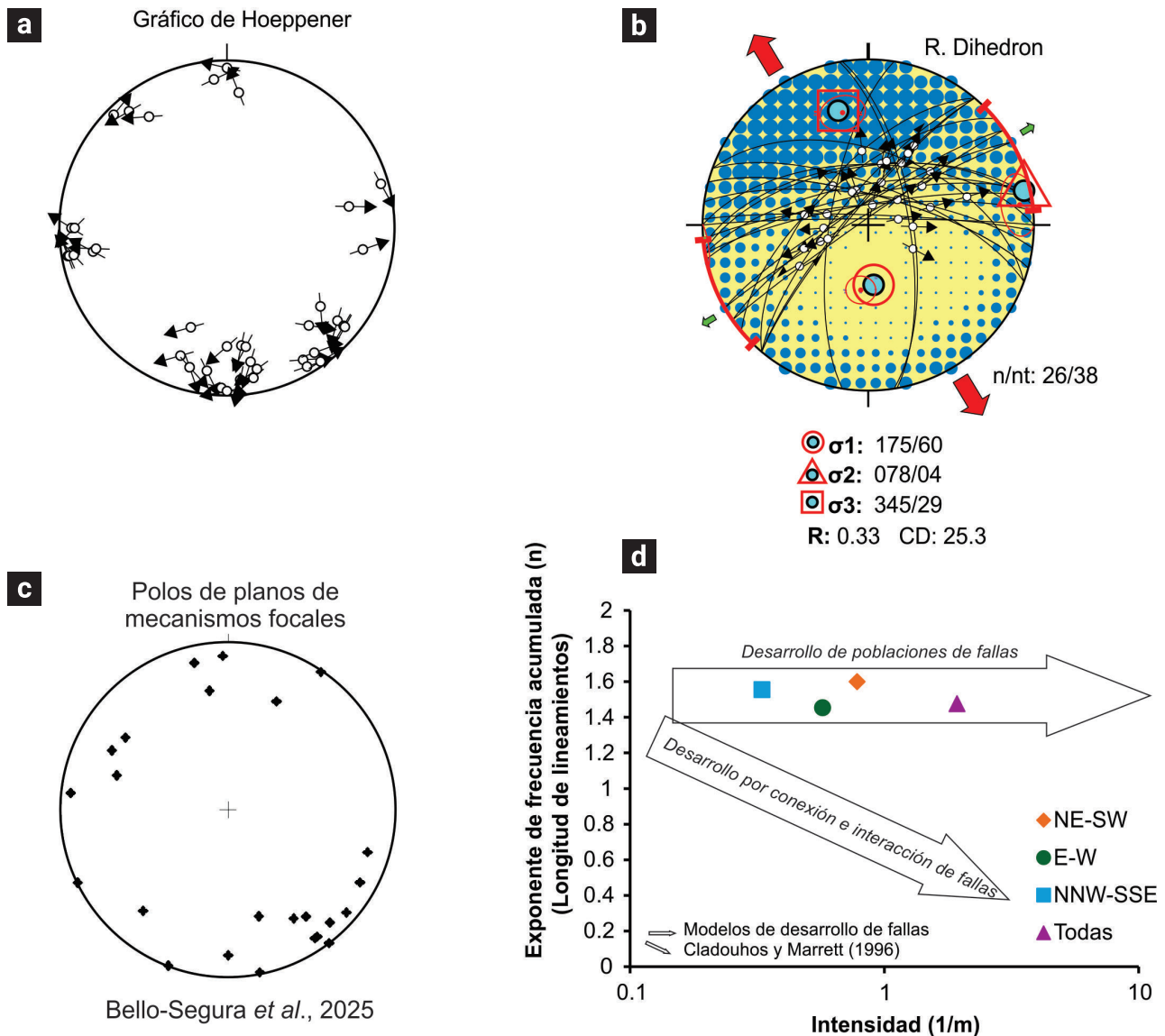


Figura 10. a, b) Gráfico de Hoeppener y solución de inversión de estrías para las fallas mesoscópicas medidas dentro de la zona de cizalla Contreras. Nótese que las direcciones de flujo de material (flechas negras en el gráfico de Hoeppener) van hacia σ_2 , y σ_3 , lo cual sugiere condiciones de deformación tridimensional (Del Pilar-Martínez *et al.*, 2020b). c) Distribución de polos de planos obtenidos a partir de mecanismos focales reportados en Bello-Segura *et al.* (2025), en donde es evidente la actividad simultánea de fallas con distinta orientación. d) Modelos de desarrollo de fallas según Cladouhos y Marrett (1996), en donde las estructuras que afectan a la Sierra de Las Cruces se forman a partir de un modelo de desarrollo de poblaciones de fallas. La dirección de los esfuerzos principales se obtuvo mediante el método *R. Dihedron* (Angelier y Mechler, 1977) en el programa WinTensor.

5. Discusión

5.1. La zona de cizalla Contreras y el sistema de fallas NE-SW en la Sierra de las Cruces

La zona de cizalla Contreras pertenece a un grupo de fallas con orientación NE-SW y está formada por un pilar tectónico (pilar Contreras) delimitado por dos fallas normales: falla Desierto de los Leones y falla Contreras (Figura 3). Las fallas NE-SW aparecen a partir del flanco sur del volcán Iturbide y se extiende hasta el volcán Zempoala en el sur (Figura 2). En la Sierra de Las Cruces, el grupo de fallas NE-SW tiene un ancho de 60 km y se correlaciona con el sistema de fallas Tenochtitlan, que se extiende desde las costas del Pacífico, en el estado de Guerrero, hasta la región de Apan, Hidalgo (De Cserna *et al.*, 1988; García-Palomo *et al.*, 2000, 2002, 2018; Vásquez-Serrano *et al.*, 2022).

Mooser (1992) y García-Palomo *et al.* (2008) definieron a la falla Contreras como un trazo continuo, de cinemática normal, que corta a rocas del Plioceno-Pleistoceno del volcán San Miguel. Según estos autores, el plano de la falla se inclina hacia el NW, con el bloque de techo desplazado en esa misma dirección. Sin embargo, en este trabajo se determinó que la falla Contreras es una estructura segmentada que corta a las rocas del volcán San Miguel (Figura 3) y que estos segmentos son cóncavos hacia el SE a escala kilométrica, lo cual sugiere que el bloque de techo se encuentra en el SE y no en el NW como se había propuesto previamente (García-Palomo *et al.*, 2008). Esta cinemática se confirma con los datos medidos en el campo, donde además se determinó que las fallas tienen una componente lateral izquierda (Figuras 3 y 9b). Así mismo, la geometría y cinemática normal de la falla Contreras es congruente con lo que sugiere Jaimes-Viera *et al.* (2025), aunque no reportan datos estructurales que demuestren dicha cinemática como se hace en este trabajo. En conjunto con la falla Contreras, existe una falla NE-SW llamada Desierto de los Leones (Figura 3). Esta falla corta rocas del flanco NW del volcán San Miguel y representa el límite noroeste del pilar tectónico Contreras (Figuras 2 y 3).

En la zona de piedemonte, hacia la prolongación noreste del pilar tectónico Contreras, existe un conjunto de fallas secundarias con orientaciones NNE-SSW y ENE-WSW que cortan una sucesión de depósitos volcánicos y volcanoclásticos del Pleistoceno (Figura 3). En conjunto, estas fallas conforman un arreglo de mosaico, el cual es característico de regiones donde interactúan fallas kilométricas con rumbo similar, debido a procesos de coalescencia (Cladouhos y Marrett, 1996; Fossen, 2016; Ramsay, 1967). En estas regiones suelen formarse estructuras de cola de caballo, asociadas con el desarrollo de la conexión entre estructuras mayores con cinemática lateral (Nevitt *et al.*, 2023; Witt *et al.*, 2012). La presencia del mosaico de fallas puede sugerir

que estas estructuras pudieron formarse en alguna etapa previa al Pleistoceno, ya que la cinemática más reciente para las fallas NE-SW es principalmente normal.

Dentro del cañón de Los Dinamos, al poniente de la Ciudad de México, se expone tanto el núcleo como la zona de daño de la falla Contreras (Figuras 3 y 4). El núcleo está constituido por material pulverizado (salbanda) de tonalidades blanquecinas y rojizas hacia el centro, y por una brecha con clastos angulosos de lavas dacíticas en los bordes. El espesor expuesto del núcleo es de 20 m. Estas características se pueden asociar con fallas con ángulos de fricción y cohesión muy bajos, cercanos a cero, lo que las hace susceptibles a desplazarse en condiciones de un esfuerzo diferencial menor en comparación con fallas cuyo núcleo está conformado por rocas más cohesivas (Alaniz-Álvarez *et al.*, 2000). Rodeando al núcleo existe una zona de daño de 1 km de espesor dominada por fracturas de cizalla, híbridas y extensionales. Las fracturas de cizalla tienen un arreglo agrupado en donde dominan las fracturas con espesor pequeño (Figura 8). Las fracturas extensionales por su parte presentan un arreglo agrupado caracterizado por dos patrones (Figura 8): individuos agrupados a escalas menores de 50 cm, probablemente generados por herencia o reactivación, y un patrón fractal a escalas mayores a 50 cm (Figura 8f). La existencia de los dos grupos de fracturas extensionales se confirma con la formación de dos pendientes en el gráfico de frecuencia acumulada contra la apertura, donde se observan dos grupos de fracturas con distinta distribución del tamaño (Figura 8b). Estos patrones en el arreglo y distribución del tamaño pueden ser el efecto de la sobreposición de fracturas, como se ha sugerido en Brogi (2011) y en Vásquez-Serrano *et al.* (2022).

Tomando en cuenta el espesor del núcleo de la falla Contreras (20 m) y el espesor de la zona de daño (1000 m), es posible conocer el valor de F_m (F_m =espesor de la zona de daño/espesor total de la zona de falla), el cual caracteriza la arquitectura de la falla y su permeabilidad a través del modelo conceptual de Caine *et al.* (1996). El valor de F_m para la falla Contreras es de 0.98, lo que sugiere un deslizamiento distribuido acomodado por fracturas de cizalla discretas y redes de fracturas, indicando que la falla actúa como conducto distribuido (Caine *et al.*, 1996).

Otro aspecto importante observado en este trabajo es la relación angular que existe entre las fracturas dentro de la zona de daño de la falla Contreras (Figuras 3 y 9b). Según el modelo de Riedel, la relación angular entre las fracturas X y T (mayor a 45 grados) indica que la zona de falla en el área de Los Dinamos se encuentra bajo un régimen de transpresión (Fossen y Cavalcante, 2017). Este régimen es posible gracias a la componente lateral que tiene la falla Contreras. En fallas con cinemática oblicua es común que se presenten sitios de transpresión y transtensión a lo largo de su traza (Fossen, 2016; Fossen y Cavalcante, 2017; Ramsay, 1967).

5.2. Deformación en el sector centro-sur de la Sierra de las Cruces

De las tres familias de fallas descritas en la Sierra de las Cruces, solo las fallas E-W son compatibles con la dirección de extensión NNE-SSW propuesta para el Cinturón Volcánico Transmexicano (Alaniz-Álvarez *et al.*, 2002; García-Palomo *et al.*, 2008; Vásquez-Serrano *et al.*, 2022). La actividad de las otras dos familias de fallas no puede explicarse mediante un modelo de neoformación bajo las condiciones de extensión actuales. En el sector centro-sur de la Sierra de las Cruces, estas tres familias de fallas se intersectan y afectan rocas que tienen un rango de edad de entre 2.29 y 0.39 Ma (Figura 2), lo que sugiere que su actividad inició en el Pleistoceno tardío. Además, la actividad sísmica registrada dentro de la Sierra de las Cruces indica que estas fallas han continuado su actividad hasta la actualidad (Aguilar-Velázquez *et al.*, 2025; Bello-Segura *et al.*, 2025; Quintanar *et al.*, 2024).

Un aspecto relevante de las tres familias de fallas es el registro de su actividad durante el Mioceno. En la cuenca de México y sus alrededores, particularmente en las regiones de Malinalco, Sierra de Guadalupe y Tizayuca-Chignahuapan, se ha documentado que estas fallas estuvieron activas durante el Mioceno medio-tardío con una cinemática lateral (Figura 11. García-Palomo *et al.*, 2000, 2018; Vásquez-Serrano *et al.*, 2022). La familia de fallas E-W, en la región de Malinalco, afecta a las rocas de la Formación Tepoztlán y registra un movimiento lateral derecho (García-Palomo *et al.*, 2000). En contraste, en depósitos volcánicos del Pleistoceno de la Sierra de Chichinautzin, estas fallas muestran una cinemática predominantemente normal (Campos-Enríquez *et al.*, 2015; Lenhardt *et al.*, 2010). Aunque, aún queda la duda sobre su cinemática actual, ya que se ha reportado actividad holocénica de tipo lateral izquierda (Norini *et al.*, 2006), la cual es congruente con los mecanismos focales de algunos sismos en la región centro-sur de la Sierra de las Cruces y en la Sierra de Chichinautzin reportados

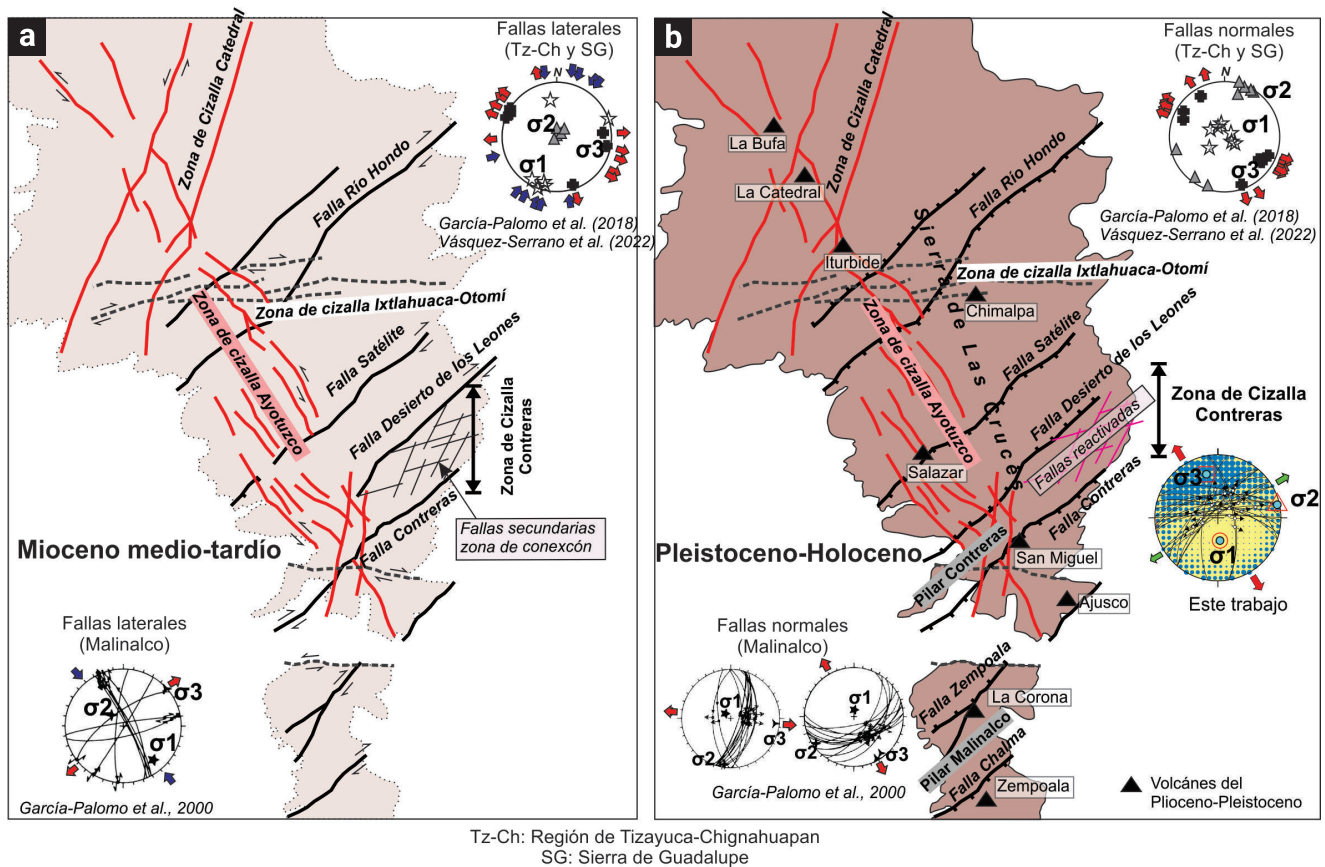


Figura 11. Modelos tectónicos para el Mioceno medio-tardío y Pleistoceno-Holoceno para la región de la Sierra de Las Cruces. a) Actividad simultánea de fallas laterales NNW-SSE, E-W y NE-SW con direcciones de acortamiento NW-SE y NE-SW. Durante el Mioceno medio-tardío, en la región nororiente de la zona de cizalla Contreras, posiblemente se forma un área de conexión entre las fallas Desierto de los Leones y Contreras, lo cual produce un arreglo en mosaico de fallas secundarias. b) Actividad simultánea de fallas normales NNW-SSE, E-W y NE-SW durante el Pleistoceno-Holoceno con direcciones de extensión NW-SE principalmente. Esta actividad se relaciona con una deformación tridimensional que reactiva estructuras previas.

por Campos-Enríquez *et al.*, (2015). Por otro lado, las fallas NNW-SSE, en la región de Malinalco y Sierra de Guadalupe, han registrado actividad durante el Mioceno medio-tardío con una cinemática lateral izquierda (García-Palomo *et al.*, 2000; Vázquez-Serrano *et al.*, 2022). Asimismo, en la región de Malinalco se ha observado una cinemática normal para estas fallas en rocas volcánicas del Plioceno-Pleistoceno (García-Palomo *et al.*, 2000). Esta última cinemática coincide con lo registrado en la región centro-sur de la Sierra de las Cruces, donde las fallas NNW-SSE son normales y forman fosas tectónicas, como en la zona de cizalla Ayotuxco (Figura 2). Esto es congruente con algunos mecanismos focales de sismos ocurridos en el flanco oriental de la Sierra de las Cruces, que sugieren igualmente una cinemática normal para estas fallas (Figura 12; Quintanar *et al.*, 2024). No obstante, en el centro de la cuenca de México, se han reportado mecanismos focales que se relacionan con fallas NNW-SSE que tienen movimiento principalmente lateral (Bello-Segura *et al.*, 2025), lo que deja abierta la interrogante

sobre cuál es su cinemática de las fallas NNW-SSE en el centro de la Sierra de las Cruces.

En cuanto a las fallas NE-SW, se ha registrado actividad de tipo lateral derecho en la Sierra de Guadalupe, y las regiones de Tizayuca-Chignahuapan y Malinalco (García-Palomo *et al.*, 2000, 2002, 2018; Vázquez-Serrano *et al.*, 2022). Este movimiento se observa en rocas del Mioceno de la formación Tepoztlán en la región de Malinalco, y en lavas andesíticas y dacíticas de la Sierra de Guadalupe (Arce *et al.*, 2019, 2020; García-Palomo *et al.*, 2000; Vázquez-Serrano *et al.*, 2022). Por otro lado, en la zona de cizalla Contreras y en otras fallas NE-SW de la Sierra de las Cruces, se observa una cinemática principalmente normal (Figuras 2, 3 y 11), la cual se restringe al Pleistoceno-Holoceno. Este movimiento es congruente con algunos mecanismos focales de sismos ocurridos en la región del piedemonte oriental de la Sierra de las Cruces (Figura 12. Bello-Segura *et al.*, 2025). Asimismo, la ubicación de los domos Cerro del Judío y Las Palmas a lo largo de la falla Contreras, así

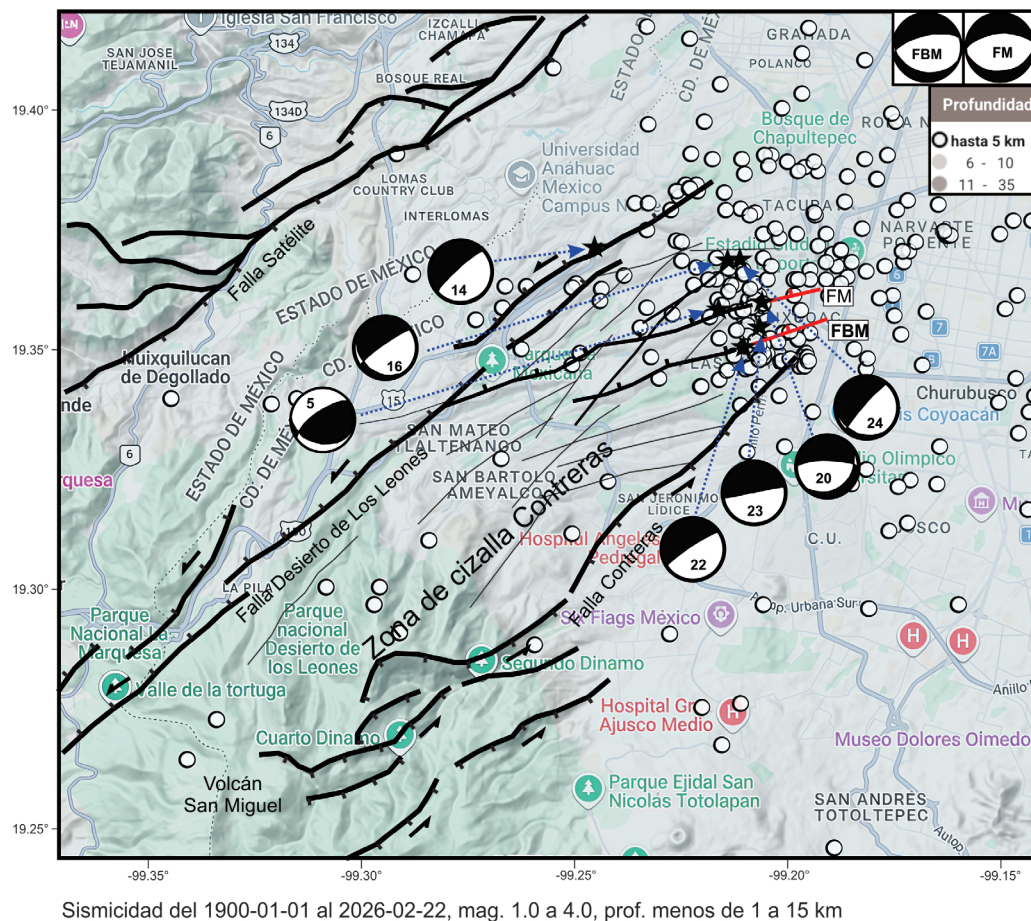


Figura 12. Actividad sísmica en la corteza superior registrada en el área de estudio desde 1900 hasta el 22 de febrero de 2026. Así mismo se incluyen algunos mecanismos focales reportados por Bello-Segura *et al.* (2025) y Aguilar-Velázquez *et al.* (2025) que tienen una cinemática similar a la zona de cizalla Contreras y se encuentran dentro de ella. La lista de los sismos utilizados para los mecanismos focales, así como la solución de dichos mecanismos se encuentran en las tablas 1, 2 y 3.

Tabla 1. Catalogo de sismos utilizados para mostrar los mecanismos focales del mapa de la figura 12 (tomados de Bello-Segura *et al.*, 2025).

Evento	Fecha	Hora (UTC)	Latitud	Longitud	Profundidad	Mw	Modelo de velocidad
5	01/12/2014	08:50:06 a. m.	19.363	-99.227	0.8	2.6	GM
14	17/07/2019	03:53:00 a. m.	19.409	-99.209	0.8	3.2	SC
16	06/07/2020	02:45:21 p. m.	19.374	-99.202	1	2.6	SC
20	11/05/2023	04:20:19 a. m.	19.364	-99.197	0.8	3.1	SC
22	12/12/2023	05:07:53 p. m.	19.363	-99.2	3	3	SC
23	14/12/2023	08:13:14 p. m.	19.363	-99.2	0.3	3.2	SC
24	14/02/2024	12:42:13 p. m.	19.374	-99.201	1.4	3.1	SC

Modelos de velocidad: GM-Modelo Genérico (modificado de Havskov y Singh, 1978); SC-Modelo Sierra de Las Cruces (Singh *et al.*, 2020).

Tabla 2. Información sobre mecanismos focales usados en figura 12 (información tomada de Bello-Segura *et al.*, 2025)

Evento	Plano 1	Plano 2	P	T
	Rumbo/echado/cabeceo	Rumbo/echado/cabeceo	Dir. Inc. /inc.	Dir. Inc./Inc
5	099/68/131	234/68/69	340/20	111/61
14	228/80/-97	82/12/-57	129/54	324/35
16	235/82/-112	127/23/-20	122/48	344/33
20	270/76/-75	042/20/-136	199/56	348/29
22	063/08/-83	236/82/-91	145/53	327/37
23	259/89/-86	004/03/-165	173/46	345/44
24	221/83/-109	111/20/-21	111/49	327/35

Tabla 3. Mecanismos focales calculados por Aguilar-Velázquez *et al.* (2025) para las fallas Barranca del Muerto (FBM) y falla Mixcoac (FM). (Ver figura 11)

Falla	Plano	Profundidad (km)	Ancho (km)	Largo (km)	Latitud	Longitud
	Rumbo/echado/cabeceo					
FBM	256.3/64.8/-105	1.18	2.6	2.6	19.369	-99.189
FM	265.1/59.2/-90.3	0.57	1.6	1.6	19.374	-99.192

como la orientación de la sierra de Santa Catarina demuestra el control estructural que tienen las fallas NE-SW en la localización y distribución de las estructuras volcánicas durante el Pleistoceno (Figuras 1 y 3).

En general, las fallas NNW-SSE, E-W y NE-SW han registrado dos periodos de actividad con distinta cinemática desde el Mioceno (Figura 11). Durante el Mioceno medio-tardío, estas fallas mostraron un movimiento principalmente lateral (Figura 11a), asociado a un régimen de acortamiento de la corteza probablemente relacionado con la dinámica de la subducción en el sector oriental del Cinturón Volcánico Transmexicano (Vásquez-Serrano *et al.*, 2022, 2025). Posteriormente, a partir del Plioceno-Pleisto-

ceno, las condiciones tectónicas cambiaron y se reactivaron las fallas NNW-SSW, NE-SW y E-W, pero ahora con una cinemática de tipo normal (Figura 11b). El movimiento simultáneo de fallas con distinta orientación y direcciones de extensión puede explicarse mediante un modelo de deformación tridimensional (Reches, 1978; Reches, 1983; Reches y Dieterich, 1983), con la reactivación de estructuras previas, como se ha propuesto para la Mesa Central en el centro de México (Del Pilar-Martínez *et al.*, 2020a, 2020b; Nieto-Samaniego y Alaniz-Álvarez, 1995, 1997). En este contexto, el esfuerzo necesario para reactivar una falla previa suele ser mucho menor que para generar una nueva (Alaniz-Álvarez *et al.*, 1998, 2002). En casos donde existe una

deformación tridimensional con la reactivación de estructuras previas, es posible observar una distribución de polos de planos de fallas que no es romboédrica (Marrett y Allmendinger, 1990; Nieto-Samaniego y Alaniz-Álvarez, 1997), como se aprecia en el diagrama de polos de planos de fallas mesoscópicas medidas en la zona de cizalla Contreras (Figura 9a). En el diagrama de Hoeppener (Figura 10a) las flechas indican un flujo de material hacia las direcciones de σ_2 y σ_3 , lo cual soporta la idea de una deformación tridimensional (Del Pilar-Martínez *et al.*, 2020b). Asimismo, la distribución de polos de planos de los mecanismos focales para la cuenca de México reportados por Bello-Segura *et al.* (2025), también sugiere actividad simultánea de fallas con distinto rumbo bajo condiciones de deformación tridimensional (Figura 10c; Reches, 1983).

Con base en los datos estructurales obtenidos en este trabajo, el eje de extensión mayor tiene una dirección NNW-SSE en la región centro-sur de la Sierra de las Cruces (Figuras 11b). Esta misma dirección de extensión, se ha reportado para la Sierra de Guadalupe, y las regiones de Tizayuca-Chignahuapan y Malinalco para el Plioceno-Pleistoceno (Figura 11; García-Palomo *et al.*, 2000; 2002; 2018; Vásquez-Serrano *et al.*, 2022). Asimismo, las direcciones de extensión obtenidas por Bello-Segura *et al.*, (2025) para algunas zonas de la cuenca de México (direcciones N-S, y NNW-SSE) son similares con las obtenidas en este trabajo.

Un aspecto interesante que ha sido poco explorado en regiones con actividad simultánea de fallas con distinta orientación es la cantidad de deformación que acomoda cada familia de fallas. En el caso de la Sierra de las Cruces, las tres familias de fallas reconocidas (NNW-SSE, E-W y NE-SW) probablemente acomodan distinta cantidad de deformación. Una primera aproximación para explorar cómo se distribuye la deformación consiste en evaluar la relación frecuencia-tamaño de las estructuras y su intensidad (Cladouhos y Marrett, 1996; Nieto-Samaniego *et al.*, 2005; Vásquez-Serrano *et al.*, 2018). Si se toma en cuenta el mapa de lineamientos de la figura 5, se puede calcular la intensidad de fracturas o discontinuidades para cada familia de fallas. La intensidad de fracturas (P21) se define como la longitud total de las trazas de fractura por unidad de área (Nieto-Samaniego *et al.*, 2005), y en este trabajo se obtuvo con la suma de las longitudes de los lineamientos entre el área que delinea la Sierra de las Cruces con base al mapa geológico de Arce *et al.* (2019) (aproximadamente 2 millones de metros cuadrados). La intensidad de fracturas puede ser considerada como una aproximación de la cantidad de deformación frágil que acomodan las rocas (Vásquez-Serrano *et al.*, 2018). Asimismo, la longitud tiene una relación de potencia con el desplazamiento que acomodan las fallas, lo cual refleja la cantidad de deformación (Cladouhos y Marrett, 1996). Las fallas NNW-SSE, E-W y NE-SW presentan valores de

intensidad de 0.33, 0.57 y 0.78, respectivamente. Estos valores sugieren que las fallas NE-SW acomodan la mayor deformación, mientras que las fallas NNW-SSE registran la menor deformación (Figura 10d). Esta distribución de la deformación apoya la idea de Alaniz-Álvarez *et al.* (1998), quienes sugieren que las fallas NNW-SSE acomodan la menor tasa de extensión en comparación con las otras familias de fallas (NE-SW y E-W). Otro aspecto interesante es la poca variación del exponente n (exponente de frecuencia acumulada) en función de la intensidad de fracturas. Dicho patrón desafía al modelo de desarrollo de fallas por crecimiento y coalescencia propuesto por Cladouhos y Marrett (1996) (Figura 10d). Este modelo plantea que las fallas se originan a partir de la formación, crecimiento y unión, lo que resulta en una disminución gradual del valor del exponente n a medida que las fallas alcanzan mayores longitudes. Los valores de n más bajos son indicativos de fallas que acomodan mayor deformación (Cladouhos y Marrett, 1996). Sin embargo, la poca variación de n observada para los tres sistemas de fallas podría explicarse mediante el modelo de desarrollo de poblaciones de fallas (Cladouhos y Marrett, 1996), que sugiere que las fallas no nacen ni mueren. Este modelo es congruente con la reactivación de estructuras previas, como es el caso de la Sierra de las Cruces. La hipótesis sobre el desarrollo de las fallas y la cantidad de deformación asociada a cada familia de fallas en la Sierra de las Cruces aún requiere ser probada con más datos estructurales, como tasas de extensión, número de fracturas a escala mesoscópica y los desplazamientos (horizontal y vertical). Por ello, resulta necesario un estudio detallado de las fallas NNW-SSE y E-W, así como de una recopilación mayor de datos estructurales a diferentes escalas dentro de la Sierra de las Cruces y en otras regiones de la cuenca de México.

5.3. Implicaciones sismotectónicas del sector oriental de la Sierra de las Cruces

El conocimiento detallado de la geometría (largo y ancho), cinemática, e historia de deformación de las fallas, así como el tipo de roca en su núcleo y medio circundante, es fundamental para conocer el potencial sísmico de una falla dentro de una región tectónicamente activa (Ovsyuchenko *et al.*, 2009). Estas variables suelen usarse en la generación de modelos de velocidad de ondas sísmicas, y la estimación del tamaño de los sismos que pueden ocurrir en una falla (Lee *et al.*, 2016; McCalpin y Nelson, 1996; Singh *et al.*, 2020). Nuestro trabajo contribuye con información importante sobre las características estructurales de la zona de cizalla Contreras, la cual registra una actividad principalmente de tipo normal durante el Pleistoceno-Holoceno. Tomando como referencia la base de datos del Servicio Sismológico Nacional (SSN) para sismos

corticales (de menos de 1 a 15 km de profundidad) registrados desde el 1 de enero de 1900, hasta el 22 de febrero del 2026, es posible observar que la región centro-sur de la Sierra de las Cruces ha registrado actividad sísmica en los últimos 100 años ([Material Suplementario 1](#) y Figura 12). Particularmente, a lo largo de la zona de cizalla Contreras, tanto en sus límites (fallas Desierto de Los Leones y Contreras), como a su interior existe actividad sísmica, sobre todo en su sector nororiental, en la zona de piedemonte (zona de Lomas) de la Sierra de las Cruces (Figura 12). Algunos de los mecanismos focales publicados por Bello-Segura *et al.* (2025) son congruentes con la cinemática de las fallas observadas en este trabajo, los cuales sugieren un movimiento de tipo normal y con un rumbo NE-SW (Tabla 1 y 2. Figura 12).

En la región de piedemonte y su transición hacia la zona lacustre existe un grupo de fallas secundarias con un arreglo de mosaico que conecta a las fallas kilométricas Desierto de Los Leones y Contreras (Figuras 3 y 12). Hipotetizamos que posiblemente sean producto de la reactivación de estructuras de conexión entre dos fallas laterales previas, activas durante el Mioceno medio-tardío como lo sugiere la cinemática reportada en rocas miocénicas (Figura 11; García-Palomo *et al.*, 2000; Vásquez Serrano *et al.*, 2022). Patrones de fallas secundarias como la observada dentro de la zona de cizalla Contreras es común en sistemas de fallas laterales (Fossen, 2016) como es el caso de la Provincia de Fallas Laterales de Chiapas (Vásquez-Serrano *et al.*, 2024; Witt *et al.*, 2012), o la zona de cizalla de la Falla de San Andrés (Nevitt *et al.*, 2023). Dentro de este arreglo de mosaico de fallas han ocurrido los llamados microsismos, los cuales se han relacionado con estructuras que tienen una cinemática principalmente normal (Figura 12. Aguilar-Velázquez *et al.*, 2025; Bello-Segura *et al.*, 2025; Quintanar *et al.*, 2024). Dos de estas estructuras secundarias (fallas de menor tamaño), dentro de la zona de cizalla Contreras, han sido caracterizadas por Aguilar-Velázquez *et al.* (2025), quienes las nombran como: falla Barranca del Muerto y falla Mixcoac (Tabla 3 y Figura 12). Estas fallas presentan un movimiento normal, congruente con fracturas mesoscópicas dentro de la zona de piedemonte observadas en este trabajo (Figura 3).

La profundidad (0.8 a 3 km) a la cual ocurren algunos microsismos (Material suplementario 1) es la misma a la que se encuentran los depósitos del Mioceno en la zona de piedemonte (Arce *et al.*, 2019), lo cual apoyaría la hipótesis de que el lugar en donde ocurre sismicidad corresponde a profundidades donde existen rocas previamente fracturadas (García-Palomo *et al.*, 2000; Vásquez Serrano *et al.*, 2022). Estudios realizados en núcleos recuperados de pozos profundos en el centro de la cuenca de México (ej. pozo Agrícola Oriental; Figura 1), han reportado zonas de falla con una orientación NE-SW y cinemática

normal a profundidades de ~1.5 km, cortando rocas del Mioceno (Vásquez-Serrano *et al.*, 2019). Los núcleos recuperados de otros pozos dentro de la cuenca de México, pueden ser la clave para entender las condiciones bajo las cuales ocurren los microsismos, debido a que son una muestra directa de las rocas en donde ocurren estos movimientos. Así mismo, se puede obtener información valiosa en vetas de calcita y cuarzo sobre variaciones en la saturación de fluido en las zonas de falla, con lo cual se puede corroborar la hipótesis sugerida por Aguilar-Velázquez *et al.* (2025) para la variación mecánica en el poniente de la ciudad de México. Es importante que en el futuro se realicen estudios detallados de las estructuras en rocas del Mioceno o más antiguas en los alrededores de la Ciudad de México y en pozos profundos, con el fin de conocer sus propiedades mecánicas y estructurales, ya que son un análogo de las rocas donde ocurren los llamados microsismos.

6. Conclusiones

La zona de cizalla Contreras al poniente de la Ciudad de México, representa una estructura que ha tenido actividad entre el Pleistoceno-Holoceno. La zona de cizalla presenta una estructura compleja, formada por dos fallas normales principales con orientación NE-SW (fallas Contreras y Desierto de Los Leones), las cuales representan los límites de un pilar tectónico que corta a rocas del volcán San Miguel. En la porción NE de la zona de cizalla existe un conjunto de fracturas, principalmente de cizalla, que forman un arreglo de mosaico. Estas fracturas tienen orientaciones NE-SW y ENE-WSW. Asimismo, dentro de la región centro-sur de la Sierra de las Cruces coexisten otras dos familias de fallas normales con orientaciones NNW-SSE y E-W, activas durante el Pleistoceno-Holoceno. El registro de actividad con cinemática lateral para las fallas NE-SW, NNW-SSE y E-W en el Mioceno medio-tardío reportada en trabajos previos, así como la actividad simultánea de estas familias de fallas durante el Plioceno-Holoceno, sugiere que la región centro-sur de la Sierra de las Cruces está afectada por deformación tridimensional, la cual reactiva a estructuras previas que probablemente son las responsables de los microsismos que han ocurrido en años recientes al poniente de la ciudad de México.

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8. References

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A preliminary approach on subway metro stations and their gravity signature

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Abstract

A specialized urban gravity survey was conducted using field measurements to determine the gravitational effects of human-made subsurface structures and, more specifically, subway stations. Large-scale underground infrastructure is common in urban environments and disturbs the gravity field. During such gravity surveys, data corrections for these types of disturbances should be calculated and applied to ensure accurate results. This study involves acquiring a set of time-lapse gravity measurements, at the location of a subway station, both before and after its construction. The gravity variation results from the removal of the subsurface geological material and its partial replacement with structural materials, as well as the void remaining and the covering of the station with loose materials. Beyond verifying the gravity anomaly caused by the metro station, the importance of the distance from the gravity station was also verified based on the field measurements, as expected. Based on the correlation between gravity variations and distance, the gravitational effect becomes negligible at approximately 95-100 meters. However, this finding requires validation through a larger dataset.

Key words: urban geophysics, underground metro station, gravity measurements, Athens basin, gravity corrections.

Resumen

Se llevó a cabo un estudio especial de gravedad urbana con el fin de determinar el efecto gravitacional de las estructuras subterráneas artificiales y, más concretamente, de las estaciones de metro, utilizando mediciones de campo. Las infraestructuras subterráneas a gran escala son habituales en las ciudades y perturban el campo gravitatorio. Durante un estudio de gravedad urbana, se deben calcular y aplicar correcciones de datos para este tipo de perturbaciones a fin de garantizar resultados precisos. Esto implica adquirir un conjunto de mediciones de gravedad a intervalos regulares, en la ubicación de una estación de metro, antes y después de su construcción. La variación de la gravedad se debe a la remoción del material geológico subterráneo y su sustitución parcial por materiales estructurales, así como al vacío restante y la cubierta de la estación con materiales ligeros. Además de verificar la anomalía gravitacional causada por la estación de metro, también se ha verificado, como era de esperar, la importancia de la distancia desde la estación gravitatoria, basándose en las mediciones de campo. A partir de la correlación entre las variaciones gravitatorias y la distancia desde la estación de metro, el efecto de la gravedad se desvanece a distancias de entre 95 y 100 metros aproximadamente, pero es necesario validarlo con un conjunto de datos más amplio.

Palabras clave: geofísica urbana, estación de metro subterránea, mediciones de gravedad, cuenca de Atenas, correcciones de gravedad.

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1. Introduction

Athens is the capital city of Greece, with a population of 3.8 million people. The city's public transport system includes the metro network, which began operating in 2000, prior to the return of the Olympic Games to their country of origin. The subway system is fundamentally composed of two lines, the Red and the Blue, which are continuously being extended with the addition of new stations. More specifically, the Blue line has undergone significant expansion towards the Piraeus area in recent years, with the addition of six new subway stations, in the areas of *Agia Varvara*, *Nikaia*, *Maniatika*, *Korydallos*, *Dimotiko Theatro* (downtown *Piraeus*) and the Port of *Piraeus* (Figure 1) (Marinos *et al.*, 2009; Schina & Charalampidou, 2014; Vogiatzis *et al.*, 2018).

The authors have previously carried out extensive gravity mapping of the Athens region (Dilalos, 2018; Dilalos & Alexopoulos, 2017; 2019a; 2019b; 2020; Dilalos *et al.*, 2018; 2019a; 2019b) in an effort to delineate the subsurface structure of such a highly urbanized area, given that other geophysical methods are often difficult to apply. In the context of this research, the authors calculated building corrections for each gravity station, in order to remove the building effect of the infrastructure. To further minimize such interference, gravity stations were primarily located in 'open' areas. During the acquisition period of these gravity measurements, between 2013-2015, the preparations for the construction of new subway stations as part of the so-called "*Piraeus extension*" of the Blue line, had already begun (Marinos *et al.*, 2009; Schina & Charalampidou, 2014; Vogiatzis *et al.*, 2018).

It is well known that removing subsurface material affects the gravity measurement of a nearby gravity station. In general, gravity and microgravity measurements have been applied in several urban areas primarily for detecting subsurface cavities (Bolaño *et al.*, 2005; Chromčák *et al.*, 2016; Jacob *et al.*, 2020; Piroddi *et al.*, 2021; Pueyo-Anchuela *et al.*, 2010; Thierry *et al.*, 2005; Tuckwell *et al.*, 2008). Taking this into account, along with the construction of the new subway stations, the authors examine, through field measurements, how a metro station disturbs the local gravity field, in other words, the impact a subway station has on nearby gravity measurements.

It should be noted that the construction of a metro station not only involves the removal of underground geological material but also replaces part of the excavated area with construction materials, mainly concrete and steel, to support the structure. The rest of the excavated area is left vacant to accommodate train and passenger movement. Most of the stations were constructed using the cut-and-cover method (apart from *Maniatika*). Part of

the excavation above the station's roof was inevitably backfilled with materials of lower density than the original geological formations. The interstation tunnels between were constructed using a tunnel-boring machine. The depths of these metro stations extend to 26-29 meters. The estimated affected subsurface volume ranges from 90,000-110,000 m³.

In a few cases in the past, the gravity method has successfully contributed to subsurface investigations prior to or during the construction of subway metro stations and lines (Égető *et al.*, 2014; Radogna *et al.*, 2002; 2004; Su *et al.*, 2021). On the other hand, buildings and infrastructure can affect nearby gravity measurements (Debeglia & Dupont, 2002; Dilalos, 2018; Dilalos *et al.*, 2018; Jacob *et al.*, 2020; Loj & Porzucek, 2019; Nowell, 1999; Yu, 2014).

2. Geological and geotechnical setting

In Figure 1, we provide a simplified geological map of the greater area (Dilalos, 2018; Dilalos *et al.*, 2019a) mostly based on the research by Papanikolaou *et al.* (2002). The geological regime across the Athens basin is quite complicated. However, not all geotectonic units of the basin appear at the investigated subway stations. Along the extension of the metro line, the Alpine basement of the allochthonous unit of "*Ypopelagoniki*" is overlain, in places, by the local allochthonous "*Athens Unit*".

The "*Ypopelagoniki*" unit is comprised mainly of Paleozoic clastic formations at its base and the overlying Triassic-Jurassic limestones. The "*Athens Unit*" is divided into two parts: the overlying limestones and the base part, known as "*Athens Schists*". The latter formation is a geological *mélange* including shales, schistic sand-gravel, breccia, meta-sandstones, phyllites, peridotites, limestones and marls.

Most of the region is covered by the Neogene and Quaternary deposits of the Athens basin. These include talus cones and scree (*Pt.sc*) or Pliocene terrestrial deposits (*Plt*) comprised of sand and clay along with Pliocene marine deposits (*Plm*) in the Piraeus peninsula, with sandstones, marls, breccioconglomerates, marly limestones with sandstone intercalations and limestones.

Data from 125 sampling boreholes executed during the preparatory phase of the subway construction (Marinos *et al.*, 2009) revealed that *Agia Varvara* station is located on terrestrial marly deposits, while *Korydallos* and *Nikaia* stations are located on Athens Schists, covered by talus cones and scree. Furthermore, *Maniatika* station has been built on ophiolites and limestones whereas *Dimotiko Theatro* station was constructed in marine marly limestones.

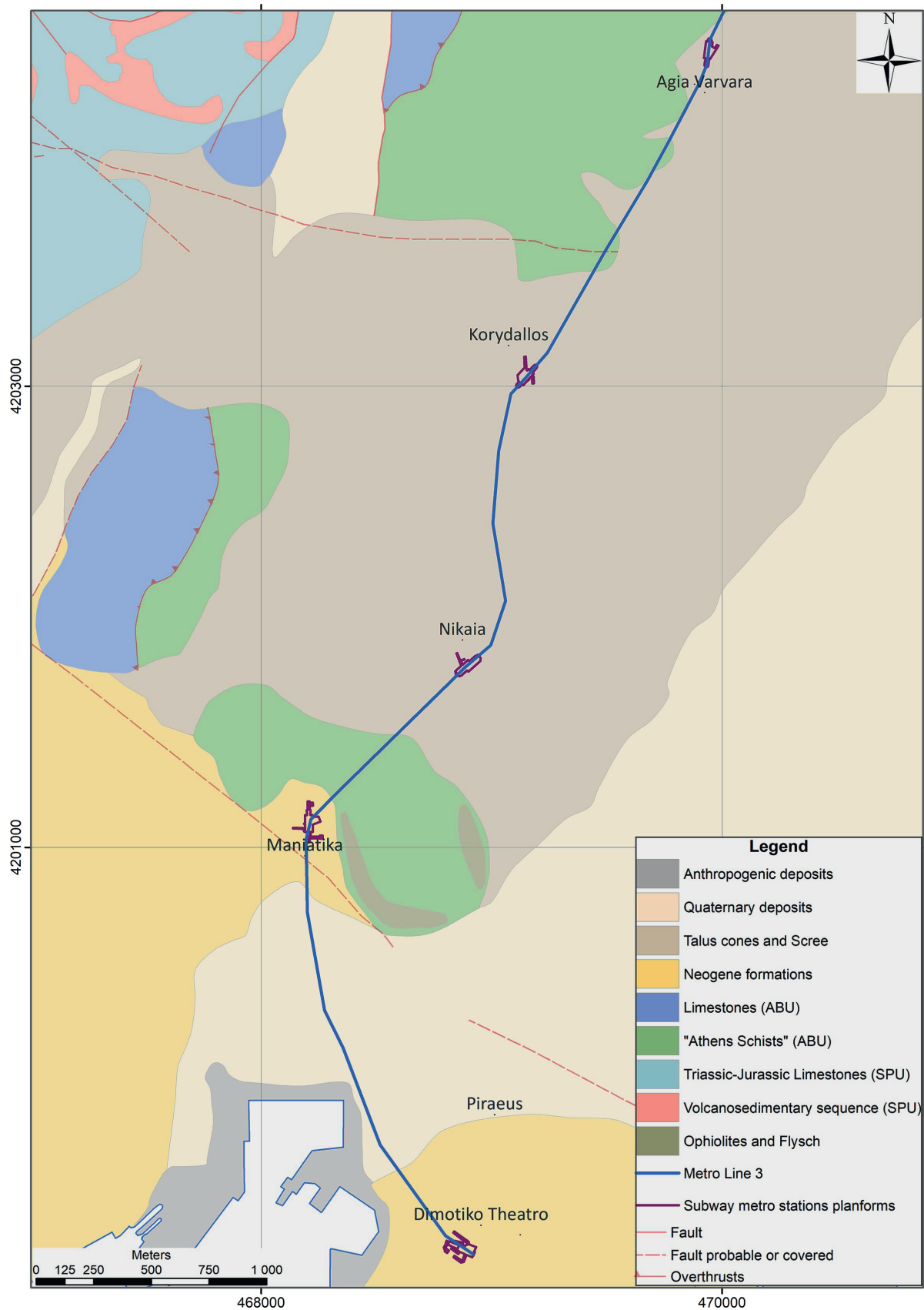


Figure 1. Simplified geological map of the area along the extension of Athens metro line 3 (blue line). The purple lines delineate the investigated metro stations platforms.

3. Gravity survey

3.1. Data acquisition

The initial gravity measurements in the areas of interest were acquired during the summer of 2013. Repeated measurements at the *Nikaia*, *Agia Varvara* and *Korydallos* stations were acquired during the summer of 2020, whereas those of the *Maniatika* and *Dimotiko Theatro* stations were acquired during the summer of 2023 due to their delayed completion.

All sets of measurements, both prior to and after construction, were acquired at the same location, each with multiple readings for verification. The locations were determined via Differential Global Positioning System (dGPS). Two individual, dual-frequency GPS antennas, were utilized, providing a vertical precision ranging from 0.6 to maximum 2.4 cm. The coordinates were determined using the Hellenic Geodetic Reference System (GGRS'87). To ensure consistency, all gravity measurements were acquired by the same person, employing the same instrument, Lacoste-Romberg G-496. Good measurement reproducibility over the years, was verified by revisiting other gravity stations around the subway stations, where no differences were expected. The *Dilaveri* gravity base, which was used for the necessary drift corrections and comparisons, is referenced to the IGSN'71 datum (Morelli *et al.*, 1974). It was established by the authors in previous works (Dilalos, 2018; Dilalos & Alexopoulos, 2019a). Furthermore, extensive photographic documentation of the surrounding area and buildings was conducted during each measurement. This allowed for the identification of any potential alterations that should be considered. Potential alterations of the gravity field included new buildings, apart from those connected to the subway station. Based on these data, no infrastructure changes were observed at the closer area of the gravity station, other than the ones directly related to the subway station.

Selecting the exact location for each gravity station proved to be challenging, as it had to meet certain criteria. For example, the elevation of the gravity station should not alter after the station's completion (e.g., due to infrastructure construction at the specific location, such as walls or decorative platforms). Additionally, we had to avoid future artificial voids at the location of the measurement, such as ventilation shafts, station exits, stairs or escalators. The location needed to be relatively quiet, situated away from traffic or operating escalators, to avoid intense oscillations of the instrument during measurements. Finally, the location of the gravity station had to remain accessible for measurements, both prior to and after the construction phase. During the initial measurements, security fences were placed around the areas that would be excavated. At this stage, the published ground plans of the subway station ("*Line 3 Extension Chaidari-Piraeus Stations*")

were immensely helpful. By consulting these plans, we could identify where planned structures that should be avoided, such as stairs, escalators, elevators, ventilation upcasts, fountains, trees and new streets. This allowed us to identify areas of significant altitude change due to artificial structures.

Moreover, the gravity measurement should preferably be acquired as close as possible to the main area lacking mass, which is practically the estimated center of the underground station. In theory, the gravity difference will be greater the closer we get to the main excavated underground volume that remains empty. Unfortunately, in some stations, such as *Agia Varvara*, safety-related accessibility restrictions prevented the gravity measurement from being taken as close to the station platform.

Taking into account all these criteria and considerations, the gravity stations were placed as illustrated in Figure 2, where the stations platforms have also been plotted, based on the georeferenced official ground plans in ArcGIS Pro.

3.2. Gravity Values

We applied only the drift and tide corrections to calculate the absolute values of each pair of gravity measurements (prior to and after construction) using Oasis Montaj software. Given that there were no changes in the terrain elevation of the closer area around the gravity stations, we did not apply terrain corrections or the Bouguer anomaly. As noted in the previous section, the gravity stations were strategically positioned to ensure that terrain corrections remain negligible. Specifically, no significant alterations were observed within the inner 20-meter segment. On the broader section of 50 meters, minor variations in structural characteristics were estimated to be less than 0.01 mGal.

The drift corrections were performed by reoccupying the *Dilaveri* reference base every couple of hours. The whole process was completed within a day to minimize the effect. Tide corrections are typically calculated with an error of ± 0.005 mGal, while the raw reading error for the specific instruments is no greater than 0.01 mGal (Dilalos & Alexopoulos, 2020). Therefore, the total estimated error is no more than 0.010-0.015 mGal. This uncertainty is small relative to most observed differences (Table 1). However, the *Agia Varvara* value is below the estimated uncertainty of the survey.

Table 1 presents the absolute gravity values of the initial and repeat measurements of all investigated locations/stations (prior to and after construction). The fourth column illustrates the gravity difference between the two measurements of each set. This reveals that the gravity field at the gravity station varied due to the construction of the corresponding metro station. As expected, the differences are negative, indicating a decrease in gravity values after the completion of the underground station.

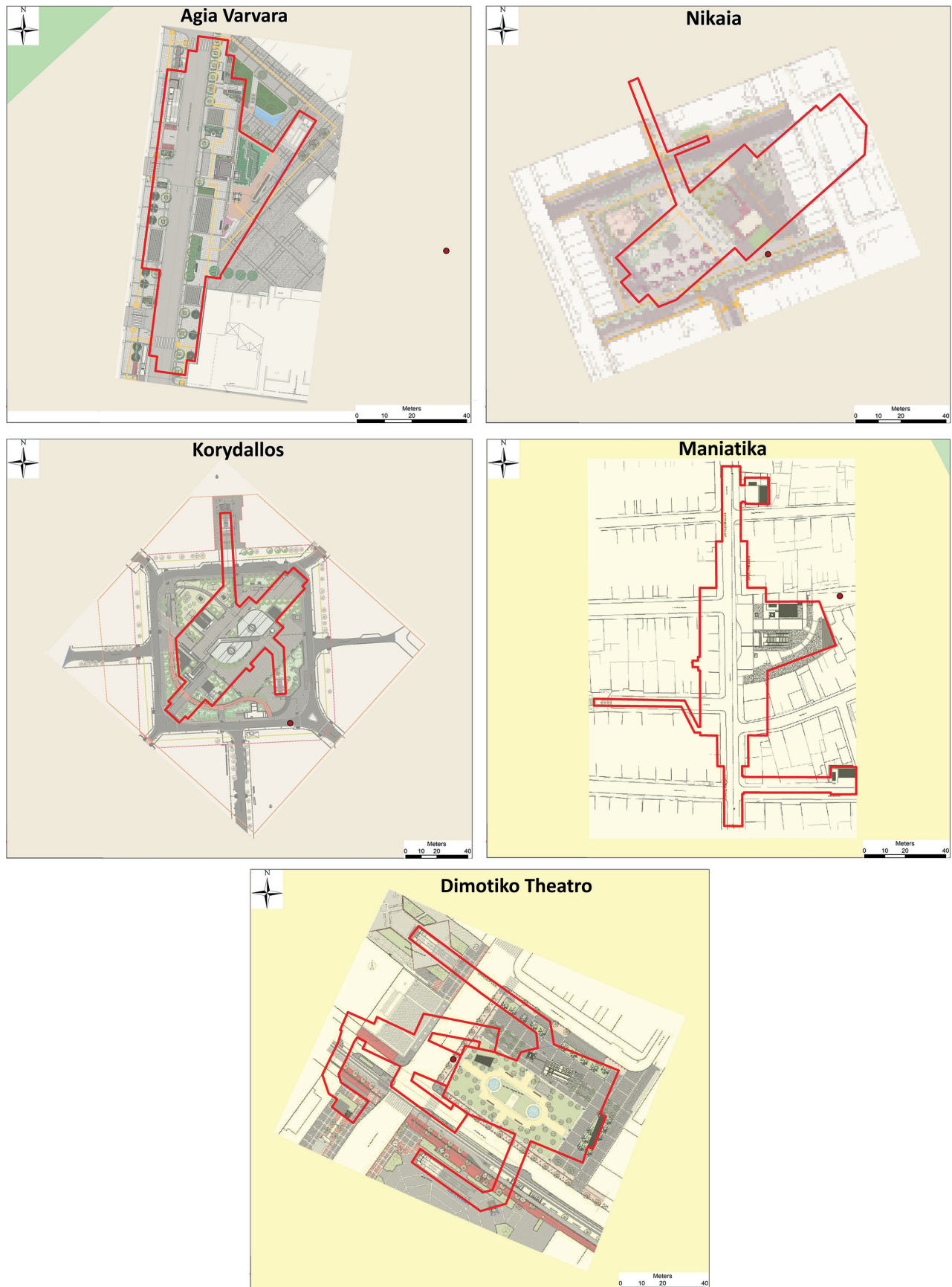


Figure 2. The stations planforms (red lines), based on the official ground plans along with the location of the gravity stations (red dots).

The average density of the subsurface volume at each station has been reduced for two reasons. Originally, the main artificial void was created due to the extended removal of (geological) mass during the construction excavations, which were partially filled in with the reinforced concrete of the station. Secondly, the constructed stations were backfilled up the original ground surface level with materials of lower density than the pre-existing geological formations.

4. Discussion

The reduction in gravitational attraction of the surface buildings is well documented (Debeglia & Dupont, 2002; Dilalos *et al.*, 2018; Loj & Porzucek, 2019; Yu, 2014), establishing that gravitational attraction weakens as the distance from a building increases. It is also known that the force of gravity decreases with the inverse square of the distance. Based on this principle, we examined the corresponding correlation of our observations, derived from the gravity field measurements prior to and after the construction of the subway stations.

The variation in gravity resulting from the construction of a subway station is also affected by the type (and density) of the geological formation in which it is built. Usually, most subway stations are built in post-Alpine formations, with relatively low densities. As previously mentioned, the *Agia Varvara* and *Dimotiko Theatro* stations are clearly located in post-alpine deposits, with calculated densities of 1.96 and 2.5 gr/cm³ respectively (Dilalos, 2018; Dilalos *et al.*, 2019a). *Korydallos* and *Nikaia* are built in the Athens Schists mélange, with a density of 2.50 gr/cm³, and the *Maniatika* station is built in alpine formations (2.67 gr/cm³). This could explain the greater reduction in gravity values at the latter station, given the greater density difference. We can adopt a density of 0.44 gr/cm³ for the subway station structure, based on the building's density, previously calculated by the authors (Dilalos *et al.*, 2018). Therefore, the density difference per station would be 1.52 gr/cm³ for *Agia*

Varvara, 2.06 gr/cm³ for *Dimotiko Theatro*, *Korydallos* and *Nikaia* and 2.23 gr/cm³ for *Maniatika* station.

Several additional factors, such as the depth of the station, the volume of subsurface excavation and the construction method, also influence gravity variations. As already discussed in the introduction section, there are no significant differences among the stations regarding these parameters.

Considering the aforementioned factors, Figure 3 illustrates the correlation between the inverse square distance ($1/r^2$) of the gravity measurement from the estimated center of the metro station and the corresponding calculated gravity variation detected after the station's completion. A logarithmic trendline has been drawn to examine the fit of the plotted data leading to the following fitting equation:

$$y = -0.115 \ln(x) - 1.089 \quad (1)$$

where y stands for the gravity variation/difference in mGal and x represents the inverse square distance ($1/r^2$) in m⁻² at which the gravity measurement was acquired from the center of the underground metro station in meters.

According to the statistical analysis of the data, the fit explains 94.9% of the variance ($R^2=0.949$). Given that only five stations contribute to the regression, the residual degrees of freedom are three, thereby forcing the t-critical value to 3.18. The logarithmic relationship proposed should be handled with caution, given the constraints of the limited dataset. Although the fit is statistically significant ($p=0.005$), the limitations imposed by the small sample size preclude the possibility of robust generalization. Consequently, the fitted equation should be regarded as a first-order empirical observation rather than a validated predictive model. The 95% confidence interval (CI) for the slope is entirely negative, ranging from -0.165 to -0.066 . Three stations fall within ± 0.014 mGal of the fitted line, which is excellent relative to the RMSE of 0.040 mGal. The gravity difference recorded at *Agia Varvara* (-0.001 mGal) falls within the total measurement uncertainty of ± 0.010 – 0.015 mGal

Table 1. The sets of gravity values, before and after the stations' construction, their gravity variation along with their estimated horizontal distance from the stations' center.

Subway station	Initial absolute gravity value (mGal)	Repeat absolute gravity value (mGal)	Gravity difference (mGal)	Distance from station's center (m)
<i>Agia Varvara</i>	980042.978	980042.977	- 0.001	95
<i>Nikaia</i>	980049.672	980049.366	- 0.306	28
<i>Korydallos</i>	980044.687	980044.569	- 0.118	70
<i>Maniatika</i>	980050.183	980049.943	- 0.240	55
<i>Dimotiko Theatro</i>	980047.492	980047.107	- 0.384	20

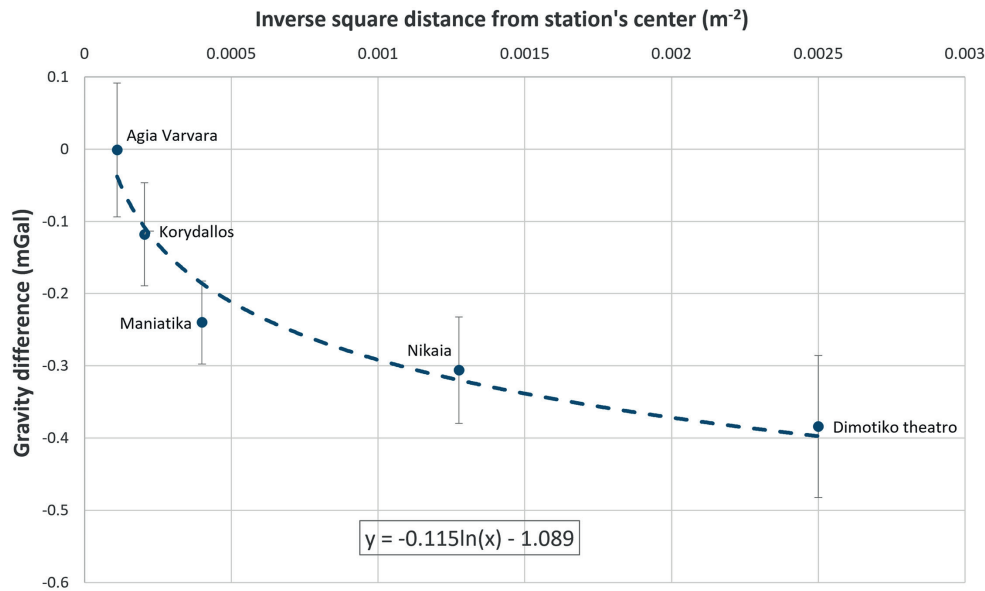


Figure 3. The correlation between the inverse square distance of the gravity measurement from the estimated center of the station with the determined gravity variation before and after its construction.

and cannot therefore be treated as a resolved gravity anomaly. *Maniatika* station is the outlier with a residual of -0.054 mGal, a value that is noticeably below the log fitting curve (Figure 3), probably because it is located in relatively denser geological formations (limestones and ophiolites). The residual could also be affected by the geometry. The error bars of Figure 3 provide a fit-confidence uncertainty only.

Based on these data, it seems that an almost negligible gravity difference will be detected at a distance of ninety-five meters from a subway station. For the stations analyzed, the results suggest that the gravity effect caused by a subway metro station becomes small at distances close to 95-100 m. Therefore, in similar urban settings, this distance may be used as a preliminary first-order indication.

It must be emphasized that there is only data from five sets of gravity measurements, one for each metro station. This does not allow us to examine the spatial distribution of the gravity variations close to the metro stations but only provides indications of it. The sample is small to make robust inferences and the results should be considered preliminary.

Finally, as a first-order verification, a simplified point-mass forward model was applied. The aforementioned representative density contrasts and distances of each station have been used along with a mean depth of 27 m and an approximate box volume of 60,000 m³. The model predicts gravity differences of 0.587, 0.378, 0.105, 0.053 and 0.017 mGal at distances of 20, 28, 55, 70 and 95 m respectively. These estimates are consistent with the observed values (Table 1) in terms of both order of magnitude

and rate of reduction, supporting the physical interpretability of the results. Discrepancies are attributed to station-specific geometry, geological variability, and the simplifications inherent in a point-mass approximation.

5. Conclusions

The impact of human-made infrastructure on gravity variation remains a relatively understudied field. The necessity of urban geophysical studies is becoming increasingly evident from several perspectives. Therefore, future urban gravity studies must fully understand the effect of surface and subsurface infrastructures on gravity data acquisition in order to remove them if necessary. Building corrections have already been studied over the last decade, as we have already mentioned, using either forward modelling or the results of field surveys. In this paper, we demonstrate the gravity effect of large-scale subsurface structures, such as the subway stations, based on field gravity measurements.

Not only have we determined the gravity disturbance of the metro stations but we have also confirmed once again that distance from the center of the gravity anomaly (here the subway station) is a crucial factor. The data suggests that, for the stations analyzed, the gravity effect becomes negligible at distances of approximately 95-100 meters. However, this estimate requires validation using a larger dataset. A similar approach and correction could be readily adopted to other large-scale subsurface infrastructure, such as underground car parks. The proposed

distance of 100 meters aligns closely with existing literature (Dilalos, 2018; Dilalos *et al.*, 2018; Debeglia & Dupont, 2002; Nowell, 1999) regarding the minimization of the gravitational attraction of buildings or basements.

Ultimately, further study is required into the spatial distribution of gravity variation resulting from such constructions. To address this, the authors have initiated a similar investigation at other metro stations under construction (Line 4), by acquiring several gravity measurements around the stations. To better understand the spatial distribution of the expected gravity differences, time-lapse gravity measurements have been planned on a grid covering the area of the future metro station. Once these stations are completed, repeat gravity measurements will be acquired to provide a more robust understanding.

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Linking Temperature, NDVI, and Methane to Assess illegal dumpsites Impacts in Puebla, Mexico

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Abstract

Assessing the environmental impact of illegal dumpsites is critical due to their harmful ecological and public health effects. In Puebla City, Mexico, illegal dumpsites pose a significant challenge to urban waste management. This study aims to evaluate the environmental impact of illegal dumpsites, focusing on pollution and environmental degradation through remote sensing techniques. Using tools like Google Earth and Google Maps Street View, we identified and mapped illegal dumpsites locations based on known problem areas. The environmental assessment concentrated on three key factors over two years (2021–2022): vegetation health, surface temperature, and methane (CH₄) emissions, with road proximity and topography added as contextual factors. Surface temperature data were derived from Landsat 7 imagery, while vegetation health was measured using the NDVI index from PlanetScope imagery. Methane emissions were mapped using Sentinel-5 (TROPOMI) satellite data processed through Google Earth Engine. The resulting maps principal component analysis (PCA) revealed three major patterns across both years: 1) High methane concentrations were found in illegal dumpsites, particularly in elevated areas. 2) There was a clear correlation between illegal dumpsites in high-temperature areas and vegetation health decline. 3) The closer a waste site was to roadways, the larger its size tended to be. These findings underscore the need for improved waste management practices to reduce environmental damage and enhance sustainability in Puebla City. Combining remote sensing techniques and multivariate analysis offers a powerful approach for monitoring and managing these environmental challenges.

Key words: Illegal dumpsites, environmental pollution, waste management, remote monitoring, ecosystem degradation.

Resumen

Evaluar el impacto ambiental de los tiraderos de Residuos Sólidos Municipales (RSM) no gestionados es crucial debido a sus efectos perjudiciales en el medio ambiente y en la salud pública. En la ciudad de Puebla, México, los tiraderos a cielo abierto representan un desafío significativo para la gestión urbana de los residuos. Este estudio tiene como objetivo evaluar el impacto ambiental de los RSM, centrándose en la contaminación y la degradación ambiental mediante técnicas de percepción remota. Utilizando herramientas como Google Earth y Google Maps Street View, se identificaron y cartografiaron las ubicaciones de los basureros en áreas problemáticas conocidas. La evaluación ambiental se enfocó en tres factores clave durante dos años (2021–2022): salud de la vegetación, temperatura superficial y emisiones de metano (CH₄), y añadió la proximidad a caminos y la topografía como factores contextuales. Los datos de temperatura superficial se obtuvieron a partir de imágenes del satélite Landsat 7, mientras que el verdor de la vegetación se midió mediante el índice espectral NDVI de imágenes de PlanetScope. Las emisiones de metano se cartografiaron a partir de datos del satélite Sentinel-5 (TROPOMI) procesados en Google Earth Engine. Mediante el análisis de componentes principales (PCA) de los mapas resultantes reveló tres patrones principales en ambos años: 1) Se observaron altas concentraciones de metano en los tiraderos, particularmente en áreas elevadas. 2) Existió una clara correlación entre los tiraderos en áreas de alta temperatura y el deterioro de la salud de la vegetación. 3) Cuanto más cerca estaba un tiradero de las carreteras, mayor era su tamaño. Estos hallazgos subrayan la necesidad de mejorar las prácticas de gestión de residuos para reducir el impacto ambiental y fomentar la sostenibilidad en la ciudad de Puebla. La combinación de técnicas de percepción remota y de análisis multivariado ofrece un enfoque robusto para monitorear y gestionar estos desafíos ambientales.

Palabras clave: Tiraderos, contaminación ambiental, gestión de residuos, monitoreo remoto, degradación del ecosistema.

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1. Introduction

Illegal dumpsites cause unprecedented damage to terrestrial and marine ecosystems worldwide. The high demand for goods and services from both industry and the population leads to increased waste generation, far exceeding nature's capacity to assimilate and degrade it (Okwaro *et al.*, 2026). Due to the high volume of illegal dumping sites, which include both organic and inorganic materials, it is challenging to implement an effective management strategy. It is estimated that a large proportion of municipal solid waste in low-income countries is still disposed of uncontrolled or open dumping sites, with rates exceeding 90%, compared to less than 5% in high-income countries (Kaza *et al.*, 2018; UNEP, 2024). This issue is compounded by the rapid increase in illegal dumpsites, which reached 2.01 billion tons in 2016 and is estimated to reach 170% of that level by 2050 (Kaza *et al.*, 2018). This severely impacts environmental conditions and public health in affected regions, making the identification and assessment of their ecological impact a crucial issue. During the decomposition of solid waste, various harmful substances are released, including greenhouse gases, dioxins, and furans (carcinogenic compounds), contributing to air, water, and soil pollution (Chandrappa and Das, 2012). These elements disproportionately affect vulnerable populations, especially when actions are not taken to identify areas with uncontrolled waste sites.

Remote sensing techniques have been widely employed to monitor the environmental conditions surrounding illegal dumpsites, as demonstrated by several studies that assess the broader environmental impact of poorly managed solid waste sites. For instance, Mahmood *et al.* (2022a) introduced geospatial indices as an alternative approach to assess the environmental impact of illegal dumpsites, focusing on satellite imagery to evaluate waste-related issues. Silvestri and Omri (2007) developed a remote sensing method to identify illegal dumpsites by analyzing the radiometric properties of surrounding vegetation. Furthermore, Arafat and Rafizul (2023) applied remote sensing tools to monitor the environmental degradation around the Rajbandh landfill in Khulna City, Bangladesh.

Several studies have specifically analyzed the impact of landfills on surrounding vegetation using satellite-based indices. Mahmood *et al.* (2022b) used satellite data to assess the bio-thermal effects in illegal dumpsites, focusing on their proximity to urban areas and the resulting environmental stress on vegetation. Similarly, Borrell *et al.* (2013) examined the territorial characteristics of illegal dumpsites in Andalucía, Spain, to identify the primary factors associated with environmental degradation, including effects on local vegetation. Land surface temperature (LST) has been another critical indicator in assessing the envi-

ronmental impact of solid waste sites. Yan *et al.* (2014) analyzed multi-temporal Landsat satellite images to monitor LST at illegal dumpsites, providing insights into how waste decomposition affects local temperature variations.

Methane formation in open dumps and landfills primarily results from the anaerobic decomposition of organic waste (e.g., food waste, vegetation, and paper) under oxygen-limited conditions. In this environment, methanogenic microorganisms break down complex organic compounds, producing biogas composed mainly of methane (approximately 50–70%) and carbon dioxide (Bogner *et al.*, 2007; Kjeldsen *et al.*, 2002). This process is strongly controlled by environmental variables. Moisture content plays a key role by enhancing microbial activity, facilitating nutrient transport, and promoting the hydrolysis of organic matter, which in turn increases methane generation rates (Kjeldsen *et al.*, 2002). Temperature is another critical factor, as it regulates microbial metabolism; methane production typically increases under mesophilic ($\approx 20\text{--}40\text{ }^{\circ}\text{C}$) and thermophilic ($\approx 50\text{--}60\text{ }^{\circ}\text{C}$) conditions, while extreme temperatures inhibit biodegradation processes (Amini *et al.*, 2013). Additionally, vegetation interacts indirectly with methane dynamics, as its presence often reflects favorable moisture and nutrient conditions. However, when landfill gas emissions are high, they can inhibit vegetation growth (Xiaoli *et al.*, 2011) by altering root systems and physiological processes in the surrounding areas (Bian *et al.*, 2014). Together, these factors control the rate and magnitude of methane emissions, which constitute a significant contribution to global greenhouse gas emissions (Bogner *et al.*, 2007). Bodah *et al.* (2022) utilized Sentinel-5P TROPOMI satellite data to study emissions of methane-related pollutants, such as NO_2 and CO, from landfills. Moreover, Wang *et al.* (2024) reported that methane emissions from landfills are globally underestimated, emphasizing the critical need for accurate satellite-based monitoring of these greenhouse gases.

In Mexico, a large amount of illegal dumpsites are disposed of in uncontrolled sites (e.g. Israde-Alcántara *et al.*, 2009; Nájera-Aguilar *et al.*, 2012). Recent estimates indicate that Mexico generates over 120,000 tons of municipal solid waste per day, with a significant proportion still disposed of in uncontrolled or inadequately managed sites (Secretaría de Medio Ambiente y Recursos Naturales [SEMARNAT], 2020; 2023). However, nearly half of these sites lack the basic infrastructure to contain pollution (SEMARNAT, 2020). This situation arises because the regulatory framework that sets out the criteria and standards for the proper management of final disposal sites is not enforced, as authorities lack the technical capacity to oversee compliance (Bernache Pérez, 2014). This issue is evident across all cities in the country, particularly in Puebla, where around 300 tons of waste from illegal dumpsites are disposed of daily at uncontrolled

sites, including ravines, rivers, streets, and vacant lots (IMPLAN, 2019). This exacerbates environmental conditions in the region and impacts public health.

The overall aim of this study is to evaluate the impact of illegal dumpsites in Puebla, Mexico; using remote sensing and geographic information systems (GIS), to provide spatially implicit evidence that can inform remediation strategies and improve solid waste practices. Hence, we ask the following questions: 1) What is the spatial distribution and density of uncontrolled illegal dumpsites in the metropolitan area of Puebla? 2) How do environmental indicators such as vegetation health, land surface temperature (LST), and atmospheric methane (CH_4) concentrations vary in areas where potential illegal dumpsites are located compared to their surroundings? 3) Can multivariate analysis (i.e., principal component analysis) effectively distinguish environmental degradation patterns associated with illegal dumpsites?

We postulate the following overall hypothesis: illegal dumpsites in Puebla City result in measurable environmental degradation, as evidenced by reduced vegetation health, increased

land surface temperatures, and elevated methane concentrations. Those impacts can be reliably detected, quantified, and spatially characterized using remote sensing and GIS assessment.

The relevance of this study is to provide a methodology applicable to other locations within Mexico or Latin America, as well as enhancing understanding of the ecological consequences of poor waste management. Also, this study supports the design of evidence-based remediation strategies and informs local policy, which can contribute to more sustainable urban environmental governance.

1.1. Study area

Puebla, the capital city of the state of Puebla in central Mexico, is situated within the Puebla-Tlaxcala Valley, bordered by the Malinche volcano to the northeast, the Popocatepetl and Iztaccíhuatl volcanoes to the northwest, and the Sierra del Tentzo to the south (Figure 1). Covering an area of 591 square kilometers, Puebla City has an estimated population of over 2.6 million, representing approximately 38% of the state's

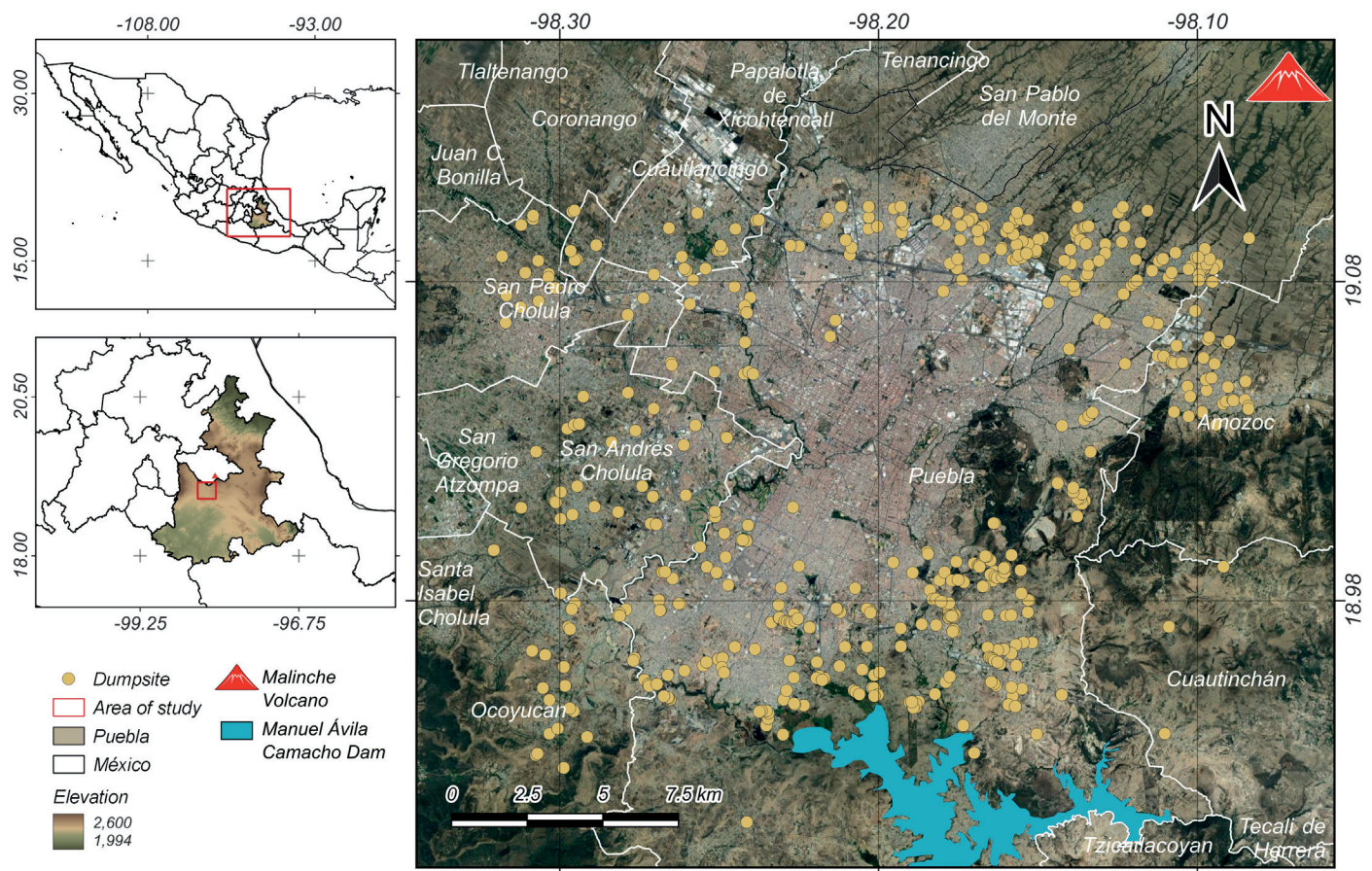


Figure 1. Location of the 502 illegal dumpsites identified in the municipality of Puebla and its metropolitan area through a visual search in Google Earth Pro.

total population (Instituto Nacional de Estadística y Geografía [INEGI], 2023). The city is intersected by various streams that feed into the Atoyac River, which flows from the northern part of the city to the south. The high population density and shifting lifestyles in Puebla have made waste management increasingly complex, leading to significant accumulation of illegal dumpsites SW in ravines, rivers, and vacant lots. This situation exacerbates environmental pollution and poses health risks to the population.

2. Methodology

This study presents a structured methodology to evaluate the environmental impact of illegal dumpsites in Puebla City. Puebla has experienced rapid population growth, leading to high population density and complex waste management challenges. We organize a workflow based on three main tasks, followed by specific steps: 1) Mapping mismanaged illegal dumpsites, and the creation of a density map to represent their spatial distribution. 2) Selection of key environmental variables related to the impact

of illegal dumpsites on the surrounding environment, such as vegetation health, surface temperature, and methane concentration. 3) Multivariate analysis to identify spatial correlation and patterns associated with illegal dumpsites. A rectangular region of interest (ROI) was defined to spatially encompass the entire Puebla metropolitan area and all identified illegal dumpsites. This extent was used to standardize methane extraction and spatial analysis across the study area. The following sections will outline the methods, tools, and techniques used for the illegal dumpsite inventory and the subsequent environmental analysis.

2.1. Inventory of solid waste sites

We conducted a systematic visual study using Google Earth Pro (images from 2019-2022), referencing known illegal dumpsites and their proximity to urban areas. We identified 502 illegal dumpsites, digitized their areas, and georeferenced them to the ITRF2008/UTM zone 14N coordinate system using QGIS. Figure 1 illustrates the spatial distribution of illegal dumpsites within the study area, while Figure 2 presents representative examples of their physical characteristics and appearance.

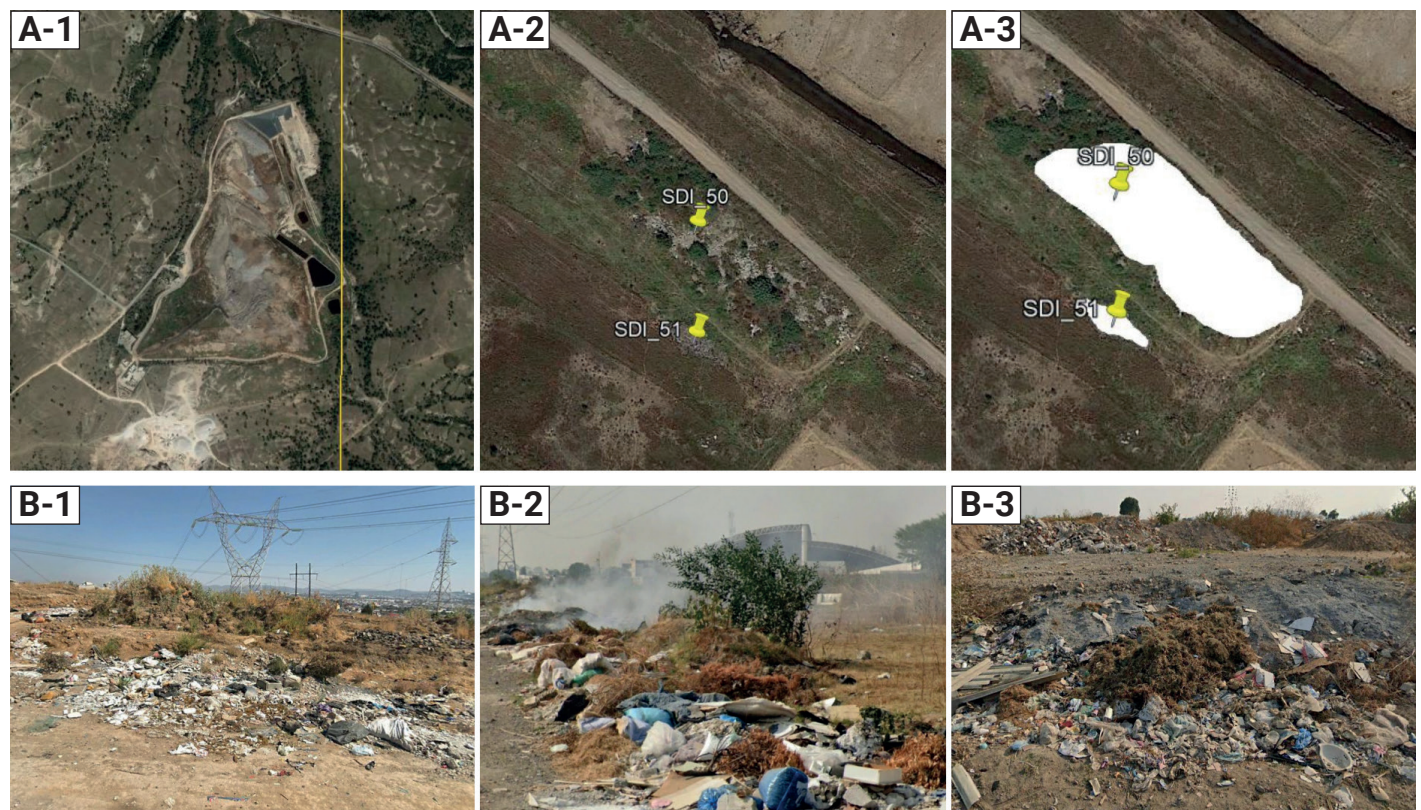


Figure 2. Representative examples of illegal dumpsites identified within the study area. (A-1) Large waste disposal site observed from satellite imagery. (A-2) Small and dispersed illegal dumpsites mapped within an urban-peripheral area. (A-3) Delineation of dumpsite areas used for spatial analysis. (B-1 to B-3) Ground-level photographs illustrating the physical characteristics, waste accumulation patterns, and environmental conditions of representative illegal dumpsites in Puebla City.

2.2. Data

In addition to the registry of illegal dumpsites, various geographic data were used to assess their environmental impact. This data includes maps of roadways and main highways, satellite imagery from the PlanetScope (2021), Landsat 7, Global Wind Atlas (GWA), and Sentinel 5P missions, and a digital elevation model retrieved from Earth Explorer. Table 1 outlines the key characteristics of these datasets, their applications in the analysis, and their sources.

2.3. Image Pre-processing

The data pre-processing aimed to correct errors and minimize atmospheric interference in Landsat 7 and PlanetScope images. Radiometric calibration was applied to adjust pixel values for sensor inaccuracies, followed by atmospheric correction to remove distortions caused by atmospheric conditions.

2.4. Inventory and Density Map

To map illegal dumpsites, a segmentation of the area of interest was established to facilitate the control of the reviewed zones. Each cell of this grid was manually examined using Google Earth and Google Maps Street View to identify solid waste sites. Additionally, images of known illegal dumpsites were used as guides for visualization and identification. A total of 502 sites were located, and their areas were calculated. Furthermore, the proximity of these sites to roadways was notorious. Once the sites were identified, a density map of these locations was created using the heatmap tool in the QGIS geoprocessing toolbox. The density map was based on the methodology of Rizzatti *et*

al. (2020), which generated a distance matrix to compute the mean and standard deviation of the distances between the sites. This allowed the definition of the influence radius to calculate the density of SW in the city.

2.5. Vegetation Greenness

A spectral index was calculated from PlanetScope images to assess vegetation greenness. The Normalized Difference Vegetation Index (NDVI), the most widely used index for vegetation assessment, uses spectral bands in the red and infrared wavelengths to distinguish between healthy and stressed vegetation, producing values that range from -1 to 1 (Huang *et al.*, 2020). Values close to zero indicate the absence of vegetation, while values near 1 indicate lush, vibrant greenery. The NDVI is calculated using the following equation 1:

$$NDVI = \frac{NIR - RED}{NIR + RED} \quad \text{Eq. 1}$$

NIR represents the reflectance in the near-infrared band, and *RED* represents the reflectance in the red band. This index provides a quantitative measure of vegetation greenness, enabling monitoring of vegetation vitality and density over time and across different geographic areas.

2.6. Land Surface Temperature (LST)

LST was calculated using Landsat 7 imagery, following the methodology established by Sobrino *et al.* (2004). This approach incorporates thermal band data and the Normalized Difference

Table 1. Description of the Imagery Used for Processing Vegetation Health, Surface Temperature, Methane Concentration, wind speed and frequency and Topographic Data for Comparison.

Image	Band	Resolution	Date	Application	Source
Planet Scope	RGB	3 m	18/02/2021 20/02/2022	Vegetation Greenness	Planet (2021)
Landsat 7	Thermal Infra-red (TIR)	60 m; 30m	20/02/2021 23/02/2022	Surface Soil Temperature; Soil Moisture	Earth Explorer
Sentinel 5p	Shortwave Infrared (SWIR)	1113.2 m	02/2021; 02/2022	Methane Concentration	Google Earth Engine
Global Wind Atlas		250 m		Wind speed and frequency	Global Wind Atlas
Digital elevation model		30 m		Topographic Features	Earth Explorer

Vegetation Index (NDVI) to accurately estimate surface temperatures. The spectral radiance is converted to brightness temperature using the following equation 2:

$$T_b = \frac{K_2}{\ln\left(\frac{k_1}{L_\lambda + 1}\right)} \quad \text{Eq. 2}$$

Donde T_b Land Surface Temperature (in Kelvin), L_λ is the Spectral radiance of the thermal band (in $\text{W/m}^2/\text{sr}/\mu\text{m}$). K_1 And K_2 are the calibration constants specific to the Landsat sensor's thermal band (666.09 and 1282.71, respectively). NDVI values are first classified into three land-cover categories to derive surface emissivity, as shown in Table 2.

After classifying the NDVI values, the percentage of vegetation P_v is calculated for each pixel in the mixed category. This is done using the following equation (3):

$$P_v = \left(\frac{NDVI - NDVI_s}{NDVI_v - NDVI_s} \right)^2 \quad \text{Eq. 3}$$

Where $NDVI_v$ represents the NDVI threshold for fully vegetated areas (typically > 0.5) and $NDVI_s$ represents the NDVI threshold for bare soil (typically < 0.2). After classifying NDVI values into vegetation, bare soil, and mixed categories, emissivity values are assigned to these land cover types. Typically, standard emissivity values are used for fully vegetated and bare soil areas, as well as for mixed land cover, and a weighted equation is applied to calculate the emissivity.

For fully vegetated areas, emissivity is set to 0.99; for bare soil, emissivity is set to 0.97. For mixed areas, where the land cover is a combination of vegetation and bare soil, the following equation is used to calculate emissivity

$$\epsilon = 0.004P_v + 0.986 \quad \text{Eq. 4}$$

Donde ϵ is the emissivity of the mixed land cover and P_v is the percentage of vegetation calculated from NDVI. The constants 0.004, and 0.986 adjust the influence of vegetation on the emissivity value.

Lastly, the surface temperature (T_s) is calculated using the following equation:

$$T_s = \frac{T_b}{1 + \frac{\lambda T_b}{\rho} \cdot \ln(\epsilon)} - 273.15 \quad \text{Eq. 5}$$

Where T_b is the brightness temperature (in Kelvin), λ is the wavelength of the emitted radiance (in micrometers, typically for Landsat 7, around 10.4–12.5 μm for thermal bands), ρ is Wien's displacement constant ($1.438 \times 10^{-2} \text{ mK}$), ϵ is the surface emissivity, which corrects for the efficiency of the surface in emitting radiation. The constant 273.15 is used to convert the temperature from Kelvin to Celsius.

2.7. Soil Moisture Mapping

The Temperature Vegetation Dryness Index (TVDI) estimates soil moisture conditions by parameterizing the relationship between land surface temperature (LST) and the Normalized Difference Vegetation Index (NDVI) (Sandholt *et al.*, 2002). The LST–NDVI feature space typically forms a triangular distribution, where the upper (dry edge) and lower (wet edge) boundaries represent limiting soil moisture conditions associated with evapotranspiration processes (Amani *et al.*, 2016, Zhao *et al.*, 2021). TVDI values range from 0 to 1, with values near 0 indicating high soil moisture and those near 1 indicating extremely dry conditions.

The Temperature Vegetation Dryness Index (TVDI) is computed as a function of land surface temperature (LST) normalized between the minimum and maximum temperature values corresponding to a given NDVI:

$$\text{TVDI} = \frac{LST - LST_{\min}}{LST_{\max} - LST_{\min}} \quad \text{Eq. 6}$$

Where LST represents the land surface temperature at each pixel, while $LST_{\min}$ y $LST_{\max}$ corresponds to the minimum and maximum temperature values associated with a given NDVI. These limits are defined by the wet and dry edges of the

Table 2. Categorization of NDVI values used for the land surface temperature methodology.

Class	NDVI
Vegetation	> 0.5
Mixed	$0.2 - 0.5$
Bare soil	< 0.2

LST–NDVI feature space and are estimated as linear functions of NDVI

$$LST_{min} = a_1 + b_1 \cdot NDVI \quad \text{Eq. 7}$$

$$LST_{max} = a_2 + b_2 \cdot NDVI \quad \text{Eq. 8}$$

Where a_1 , a_2 are the intercepts and b_1 , b_2 are the slopes of the wet and dry edge models, respectively, obtained through ordinary least squares regression.

2.8. Methane concentration map

The troposphere's methane (CH_4) concentration was estimated using data from the Sentinel-5P satellite, specifically the Tropospheric Monitoring Instrument (TROPOMI), following methodologies already known and calibrated in global and regional works (e.g., Hu *et al.*, 2018; Lorente *et al.*, 2021; Plant *et al.*, 2022; Scheneising *et al.*, 2019). According to Richa-Singh (2023), TROPOMI detects methane by measuring characteristic absorption features in the infrared region of the electromagnetic spectrum. The sensor records the intensity of solar radiation reflected from the Earth's surface and atmosphere, where methane molecules absorb specific wavelengths of light. These absorption signatures are processed through atmospheric correction and retrieval algorithms to estimate atmospheric methane concentrations spatially.

The analysis was conducted on the Google Earth Engine (GEE) platform. Data were obtained from the Sentinel-5P image collection, which provides column-averaged methane volume mixing ratios. A temporal filter was applied from January 1, 2021, to December 31, 2022, to compute the average methane concentration for the study area during this period. A rectangular region of interest was defined, and the CH_4 data were filtered to include only the observations within this area. Pixels were further filtered based on data quality, selecting only those with a quality assurance value greater than 0.5 to ensure measurements reliability. A temporal average of the filtered images was then calculated to create a representative map of methane concentration for 2021 and 2022. Finally, the mean methane concentration was visualized using a color palette, with values ranging from 1750 to 1900 parts per million to highlight variations, and the resulting map was exported as a GeoTIFF image for further analysis and interpretation.

2.9. Wind speed and frequency

The speed and frequency of the wind were obtained from the Global Wind Atlas (GWA) dataset, which provides estimates of average wind speed over the entire land surface at a spatial resolution of 250 m. The methodology used in GWA to reduce the

spatial resolution of the wind is based on linearized flow models and high-resolution topographic data applied to databases and high-resolution wind maps (Davis *et al.*, 2023). To download the data, 12 wind sampling points were spatially distributed across the study area to provide representative coverage and achieve a balanced spatial representation of local wind conditions. Information on average wind directions and frequencies was obtained in csv format. In addition, average wind speeds at a height of 100 m for the study area were obtained as a TIFF file.

2.10. Multivariate data analysis

PCA is a multivariate data analysis technique that aims to reduce the dimensionality of a correlated dataset while preserving as much variance as possible. The procedure involves computing the covariance matrix of the original variables, followed by extracting the eigenvalues and eigenvectors. The eigenvectors, which represent the directions of maximum variance in the data, are then ordered by their corresponding eigenvalues. This allows the identification of principal components and linear combinations of the original variables. By projecting the data onto these principal components, researchers can effectively identify patterns, relationships, and underlying structures within the dataset, facilitating a clearer understanding of the data's inherent characteristics. Data processing using the PCA technique was performed with the help of R software and its graphical interface Rstudio.

3. Results and discussion

This section presents the findings from the analysis of vegetation health, surface temperature, methane concentration, and the influence of topography and roadways in the study area. Additionally, the cartography of illegal dumpsites was developed to serve as a comparative factor against all the analyzed variables. The results highlight the complex interactions among these factors and their implications for environmental monitoring and management. By integrating remote sensing data with multivariate analysis techniques, we aim to uncover patterns and relationships that enhance our understanding of the spatial and temporal dynamics affecting vegetation health and land surface temperature. The insights gained from this analysis are critical for informing mitigation strategies and improving land-use planning in response to environmental changes.

3.1. MSWD Density Map

The density map (Figure 3) illustrates the areas with the highest concentrations of illegal sites in Puebla, highlighted in red in

the northeastern and southern regions. The city's southern part notably contains more illegal dumpsites. A total of 502 sites were identified, exhibiting significant variation in size. The median area of these sites is 154 m², with 75% covering less than 400 m². Most illegal dumpsites exceed 1,000 m², with the largest reaching 5,047.06 m². The identified solid waste sites also exhibit specific characteristics concerning their distance from major roadways and highways in the city. The median distance from these sites to the nearest roadways is 29 m, with 75% located within 54.4 m of a roadway. Several sites are more than 250 m away from roadways, with the maximum distance recorded at 2,221.4 m.

3.2. Vegetation Greenness

The assessment of vegetation health in the study area for 2021 and 2022, as indicated by the NDVI (Figure 4), reveals three prominent peaks at 0, 0.4, and 0.85 (Figure 3, C1 and

C2). The first peak indicates the presence of water bodies in the region. The second peak reflects moderately healthy vegetation interspersed with bare soils typical of urban environments. The final peak signifies agricultural lands and the presence of aquatic lilies. The average NDVI values for these years were 0.46 and 0.50, indicating a slight improvement in vegetation health in 2022. The boards C1 and C2 (Figure 4) show the identified peaks for 2021 and 2022, while board D (Figure 4) shows box plots summarizing NDVI statistics for each site in both years. It can be seen that for 2021, most sites have stressed and moderately healthy vegetation, with values ranging from 0.41 to 0.53, and that, for 2022, the overall condition of the vegetation improves compared to the previous year. However, despite the overall improvement, vegetation health remains affected.

The average vegetation health in the sites identified with SW shows slight variation between the two years. However, in 2021, the impact on vegetation health was more pronounced.

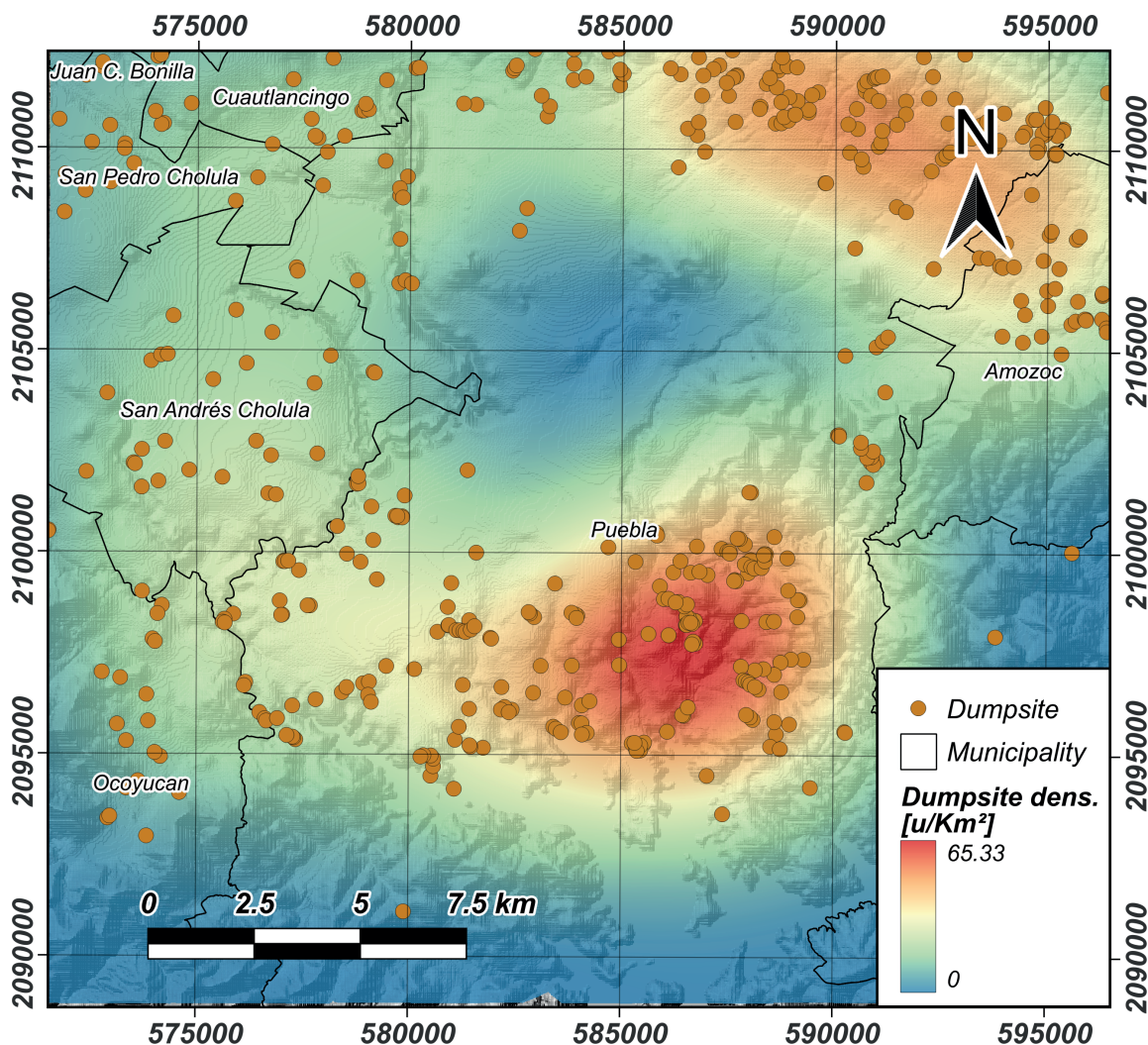


Figure 3. Density map of poorly managed SW sites identified through remote sensing.

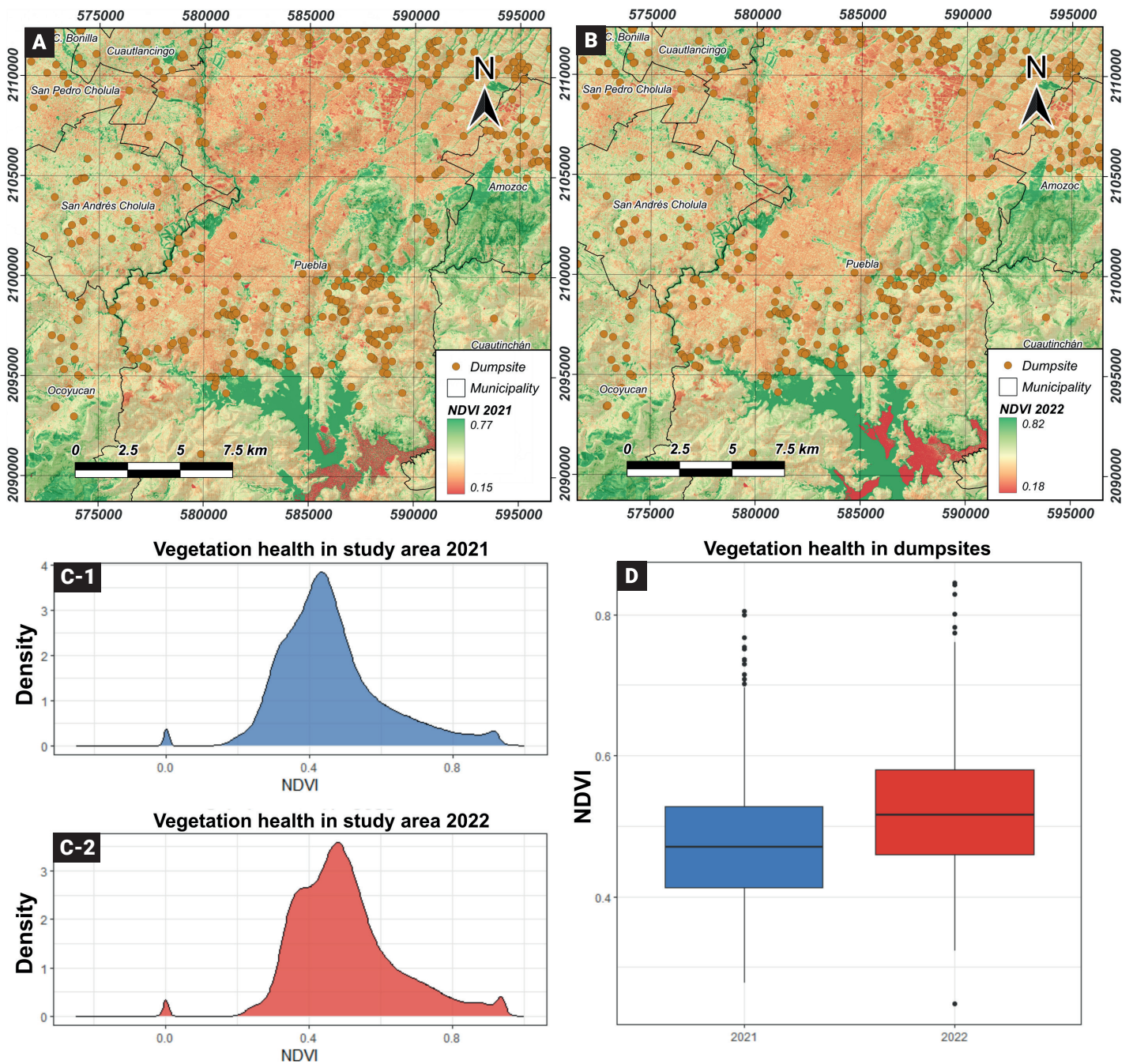


Figure 4. NDVI Analysis of Vegetation Health in the Study Area for 2021 and 2022. Panels A and B show NDVI maps, while Panels C1 and C2 illustrate the corresponding peak values. Panel D presents box plots summarizing NDVI statistics across sites for both years.

In 2022, although the minimum NDVI decreased from 0.28 to 0.25, overall vegetation improved. In addition, the first quartile limit increased from 0.41 in 2021 to 0.46 in 2022, suggesting a slight improvement in vegetation health at SW sites, although some areas remain affected.

The outliers observed in both years indicate healthy vegetation within the MSWD sites (Figure 4-D), with a general improvement in 2022. Healthy vegetation is linked to improved overall

conditions in the study area, primarily due to increased rainfall. Nevertheless, the reduction in the minimum NDVI values at the SW sites indicates ongoing issues with vegetation health. This suggests that climatic factors, which have positively influenced overall vegetation conditions, may be masking the deterioration experienced by most SW sites. Therefore, despite the overall improvement, SW continues to negatively impact vegetation health in the area.

3.3. LST

The analysis of surface soil temperature in the study area for 2021 and 2022 reveals significant trends, as shown in Figure 4. The surface temperature maps for 2021 and 2022 are shown in panels A and B of Figure 5, respectively. Specifically, the average temperature at illegal sites in 2021 was 21.14 °C, exceeding the regional average. The recorded surface temperatures show peak averages of approximately 20.5 °C in 2021 and 24.05 °C in 2022, indicating a general increase in temperature from one

year to the following (Figure 5 C1 and C2). The third-quartile temperature in 2021 was 22.37 °C, with a maximum recorded temperature of 27.60 °C. These results indicate that at least 25% of the monitored sites exhibited surface temperatures above the regional average, likely due to high soil reflectance resulting from vegetation degradation and the thermal output of ongoing organic matter decomposition. (Figure 5D; Mahmood *et al.*, 2016).

In 2022, surface soil temperatures rose at the SW sites, mirroring a similar trend across the entire study area (Figure 5B). The average temperature at these sites was 23.99 °C, which remains

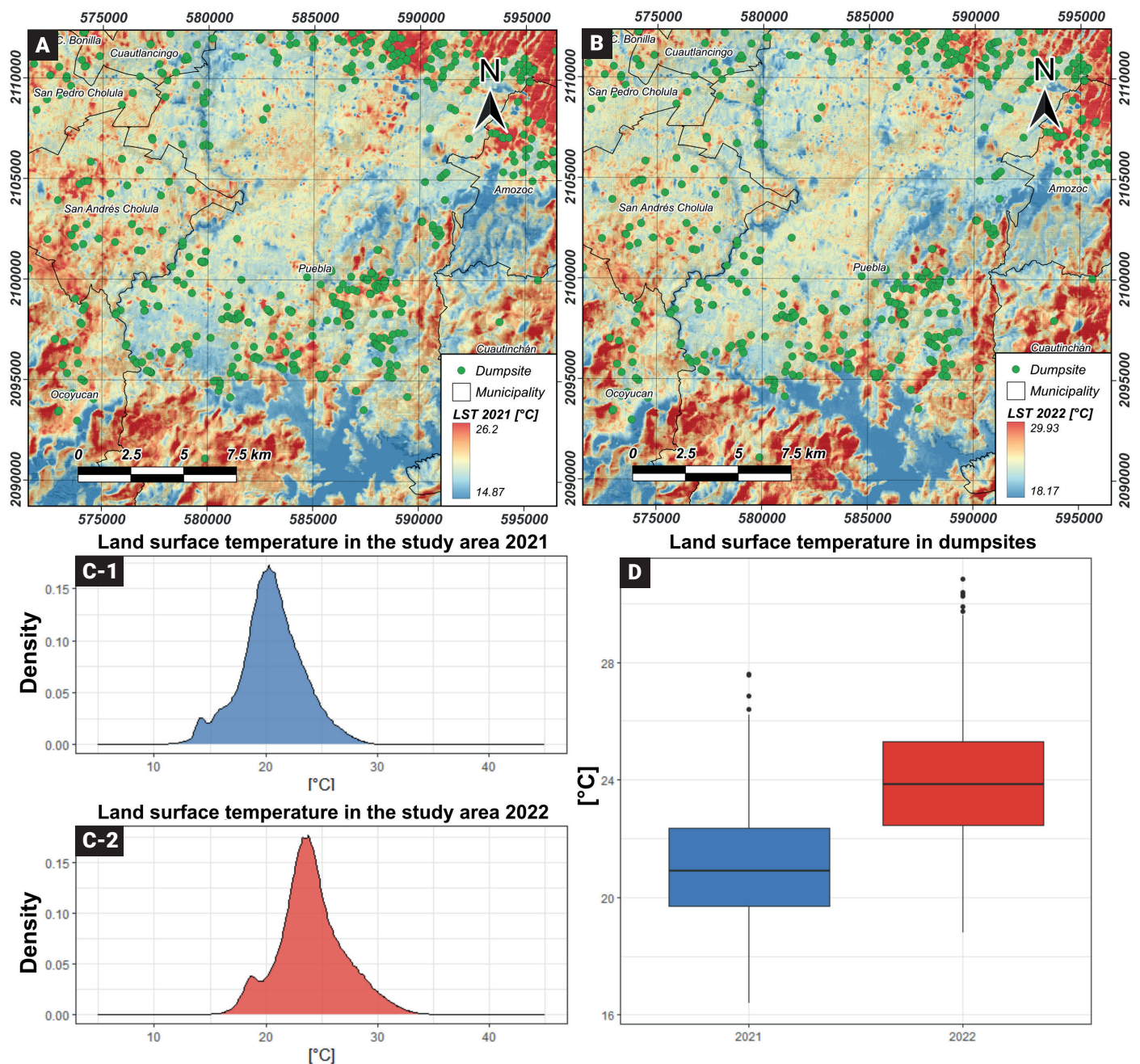


Figure 5. Surface Temperature Analysis in the Study Area for 2021 (A) and 2022 (B). C1 and C2 present the corresponding density histograms for each year. D provides box plots summarizing temperature statistics across sites for both years.

below the overall 2022 average. The third quartile ranged from 25.31 to 30.83 °C (Figure 5D). These trends are supported by the density histograms in Figure 5C1 and 5C2, which depict the temperature distributions for 2021 and 2022, respectively. The box plots in Figure 5D summarize the temperature results by site for both years, illustrating the variability and central tendency of the data. Overall, the data indicates that while temperatures have increased, the influence of illegal dumpsites remains critical to the area's thermal dynamics.

3.4. Soil Moisture Mapping

The LST–NDVI feature space for February 2021 and 2022 is shown in Figure 6. Linear relationships defining the wet edge and dry edge were derived for each period using ordinary least squares regression (Table 3).

In 2021, the wet edge is described by the equation $LST = 8.975 + 7.793 \cdot NDVI$ (Table 3), exhibiting a positive slope, indicating an increase in land surface temperature across vegetation cover under maximum evapotranspiration conditions. This positive trend may be influenced by the presence of unmasked water pixels within the study area. The dry edge is defined by $LST = 32.614 - 16.312 \cdot NDVI$, where the negative slope reflects a decrease in maximum surface temperature with increasing vegetation cover under minimal evapotranspiration conditions. This upper boundary represents the limit beyond which soil moisture is insufficient to sustain continuous evaporative flux.

In 2022, the wet edge is represented by $LST = 10.400 + 9.204 \cdot NDVI$ (Table 3), with a positive slope indicating that surface temperature increases with higher NDVI values under wet conditions. The dry edge is defined by $LST = 39.776 - 22.646 \cdot NDVI$, with a negative slope consistent with the 2021

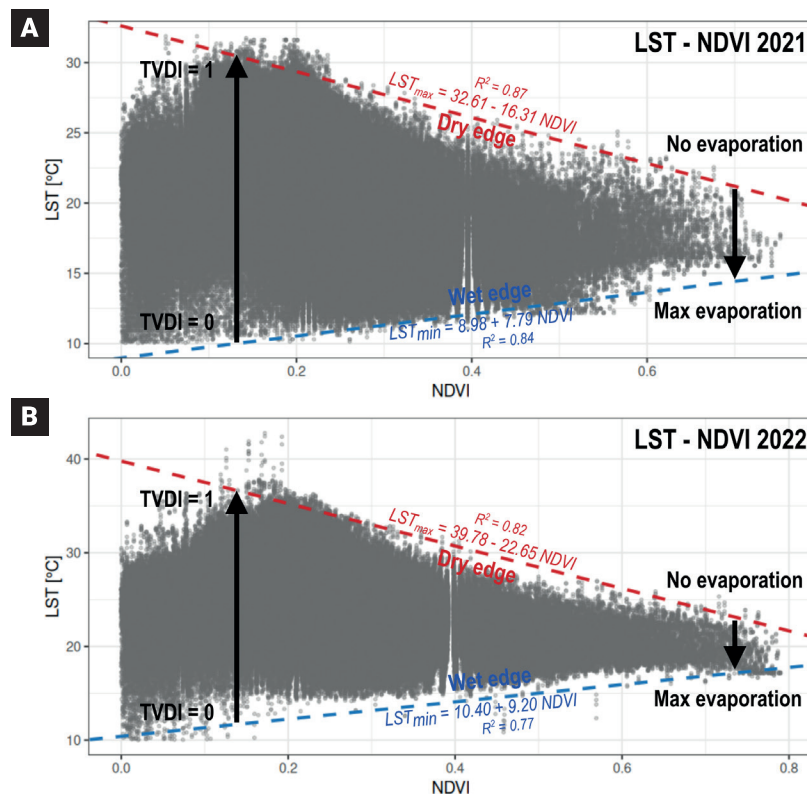


Figure 6. LST–NDVI Feature Space and Derived Dry and Wet Edge Relationships for Puebla City. A) 2021; B) 2022.

Table 3. Linear equations of the dry and wet edges derived from the LST–NDVI feature space, including their coefficients of determination (R^2) for February 2021 and 2022.

Year	Dry edge	Wet edge
2021	$32.614 - 16.312 (NDVI)$ $R^2 = 0.867$	$8.975 + 7.793 (NDVI)$ $R^2 = 0.843$
2022	$39.776 - 22.646 (NDVI)$ $R^2 = 0.819$	$10.40 + 9.204 (NDVI)$ $R^2 = 0.766$

pattern, indicating a reduction in maximum surface temperature as vegetation cover increases. Figure 6B illustrates the spatial distribution of the LST–NDVI feature space and the corresponding wet and dry edge equations for Puebla City for both periods (A: 2021; B: 2022).

The results indicate that the southwestern, eastern, and north-central areas (aquamarine colors) exhibit the highest soil moisture, whereas a large portion of the study area is characterized by high dryness. In 2021, the TVDI shows a mean value of 0.54 with a standard deviation of 0.13, indicating moderate

variability around the mean. The corresponding histogram (Figure 7C-1) displays an approximately normal distribution, with a higher concentration of values above 0.5.

In 2022, a slight decrease in the mean TVDI (0.52) is observed, along with a standard deviation of 0.12, suggesting a similar level of dispersion but a general shift toward lower values. This pattern is reflected in the histogram (Figure 7C-2), where a reduction in values above 0.5 can be observed.

TVDI values at sites with solid waste indicate relatively higher soil moisture conditions in 2022 compared to 2021. In 2021,

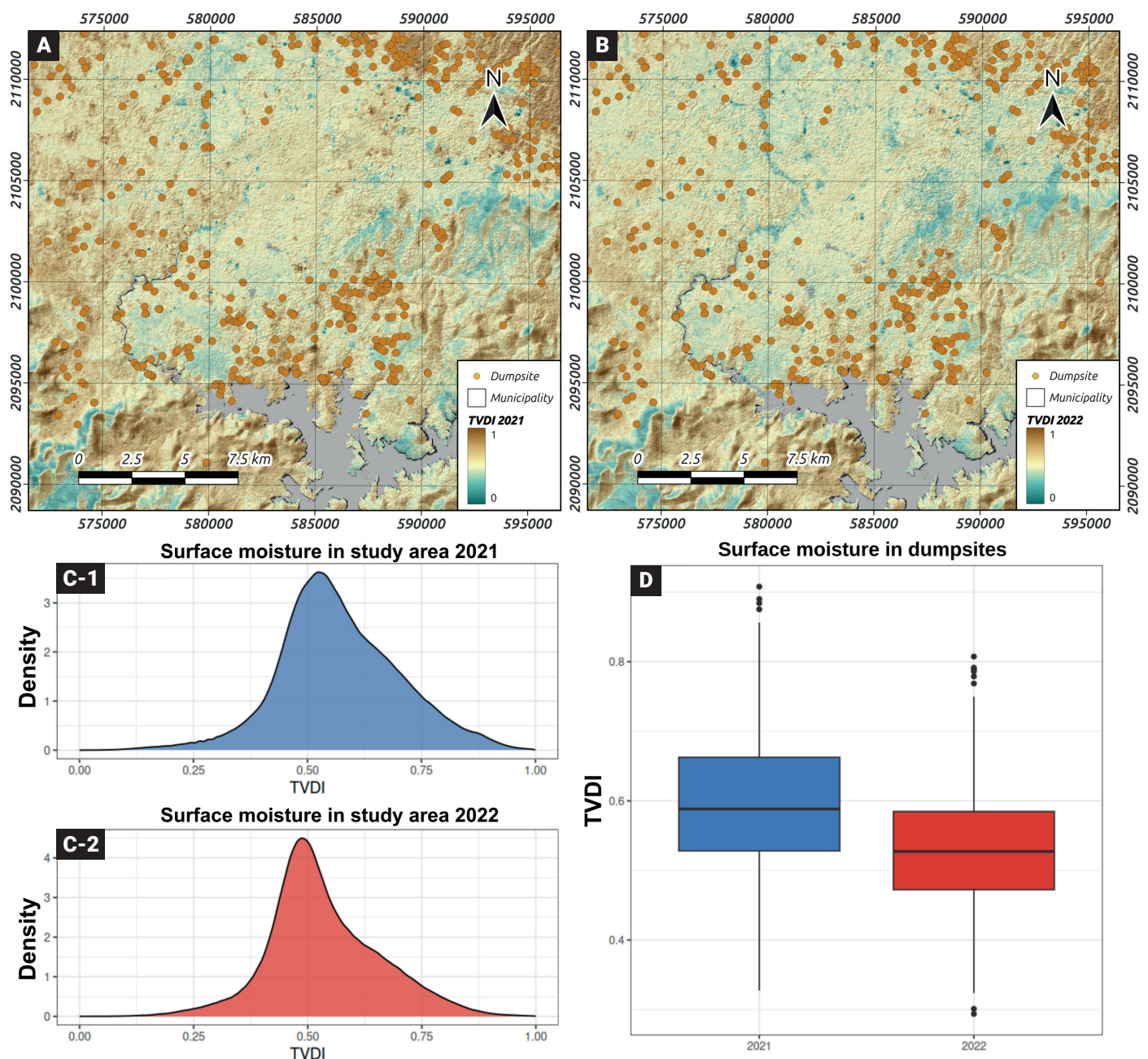


Figure 7. Soil moisture analysis for the study area in 2021 and 2022. (A–B) Spatial distribution maps of TVDI-derived soil moisture conditions for each period. (C1–C2) Corresponding density histograms. (D) Box plots summarizing soil moisture statistics across solid waste sites for both years.

most waste sites exhibit TVDI values above the median (0.55), whereas in 2022, the median decreases to approximately 0.51, and the third quartile falls below the median observed in 2021, suggesting an overall shift toward wetter conditions.

3.5. Methane concentration map

Methane concentration data were recorded in 2021 and 2022, showing notable differences between the two years. The results of methane concentration measurements in Puebla are

summarized in Figure 8. The highest levels of methane gas are found in the northeastern and western parts of the city, with elevated concentrations also detected in the central area (Figure 8A and B). Conversely, the lowest concentrations are primarily located in the southern region. Moderate methane levels are dispersed throughout the city, with lower levels observed in the central-northern area; however, this distribution does not match the concentration patterns observed in the northeast and south. Density plots indicate that the distribution of methane concentrations exhibited several peaks, reflecting significant

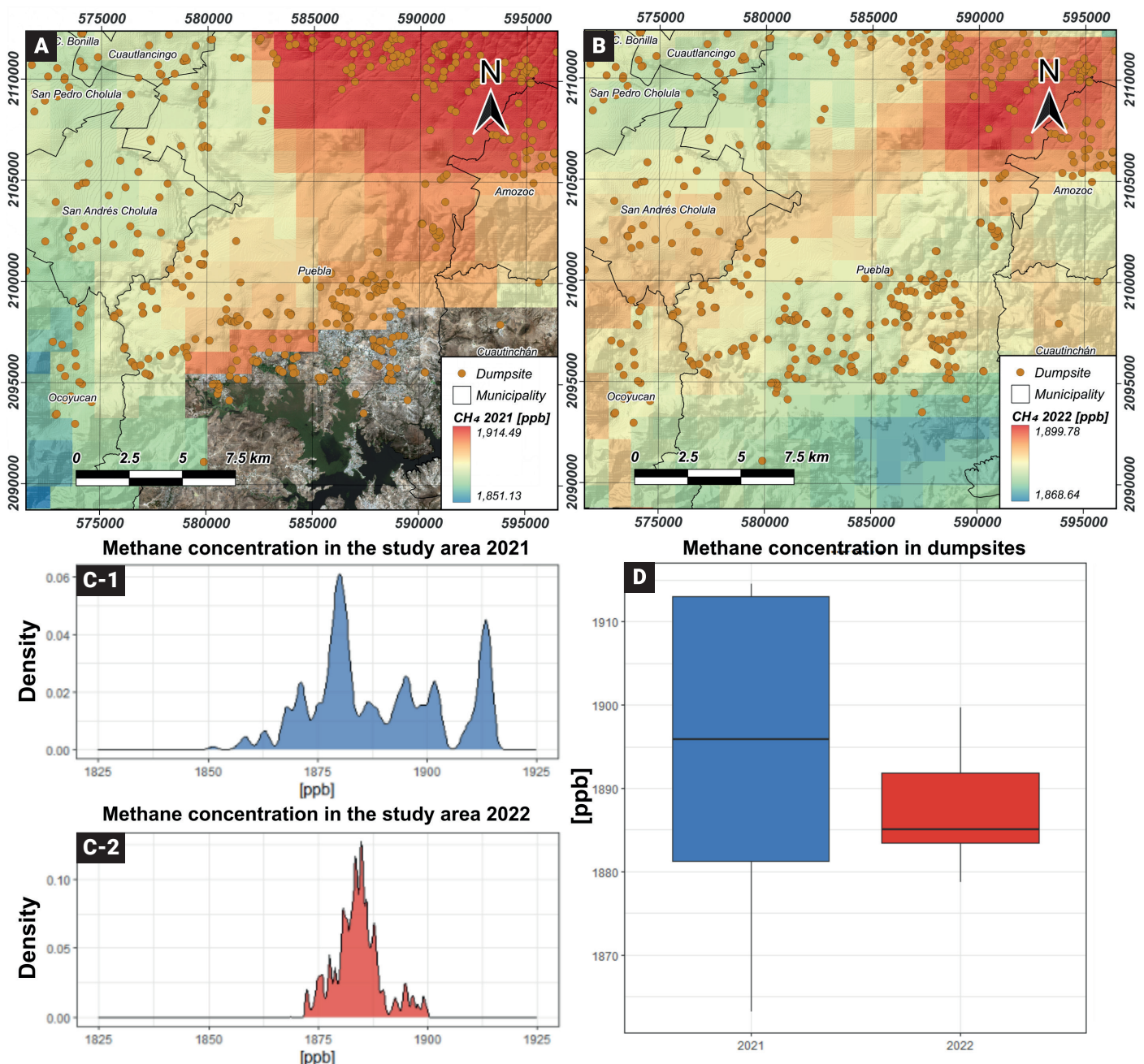


Figure 8. Methane Concentration Mapping in Puebla for 2021 and 2022. Panels A and B display concentration maps, while Panels C1 and C2 illustrate yearly density distributions. Panel D presents box plots summarizing methane concentration statistics across sites for both years.

variability in the data. In 2021, most values clustered around the mean of 1889 ppb, with high dispersion and a variance of 229 (Figure 8C1). In contrast, the 2022 methane concentration had a mean of 1884 ppb, with reduced dispersion compared to 2021, as indicated by a much lower variance of 30 (Figure 8C2). Similar variability in methane fluxes has been reported in other illegal dumpsites due to fluctuating biological activity and waste age (Zhang *et al.*, 2023 reported 750 parts per billion (ppb) to approximately 1850 ppb in 2017; Wang *et al.*, 2024 reported values around 678 ppb, generally in the USA).

For sites with illegal dumpsites, methane measurements in 2021 showed greater dispersion, with an interquartile range of 35 and a maximum of 1914 ppb (Figure 8D). This could be associated with irregular methane releases from the illegal sites decomposition. In 2022, the dispersion of measurements decreased, with minimum and maximum levels of 1875 and 1900 ppb, respectively, reflecting lower variability than in 2021 (Figure 8D).

3.6. Distribution of wind speeds and directions

The average wind speed and frequency in the study area indicate two spatial patterns. The first pattern is observed in the northeast, located on the southern slope of the Malinche volcano, where the highest methane measurements are recorded. This zone has the lowest average wind speeds in the area, ranging from 2.83 m/s to 3.34 m/s, corresponding to points 2, 5, 6, and 7 (Figures 9 and 10).

In these points, the wind exhibits a more or less regular frequency for various directions, however, there is a slightly greater predominance towards the southeast and south (Figure 10), and a lower frequency towards the southwest (Figure 10). The second pattern observed in the study area is mainly found in the southern and western parts of the city, located above the municipality of San Andrés Cholula and the Manuel Ávila Camacho dam (Figure 9). In this area, average wind speeds of 3.53 m/s to 5.19 m/s are recorded at points 1, 3, 4, 8, 9, 11, and

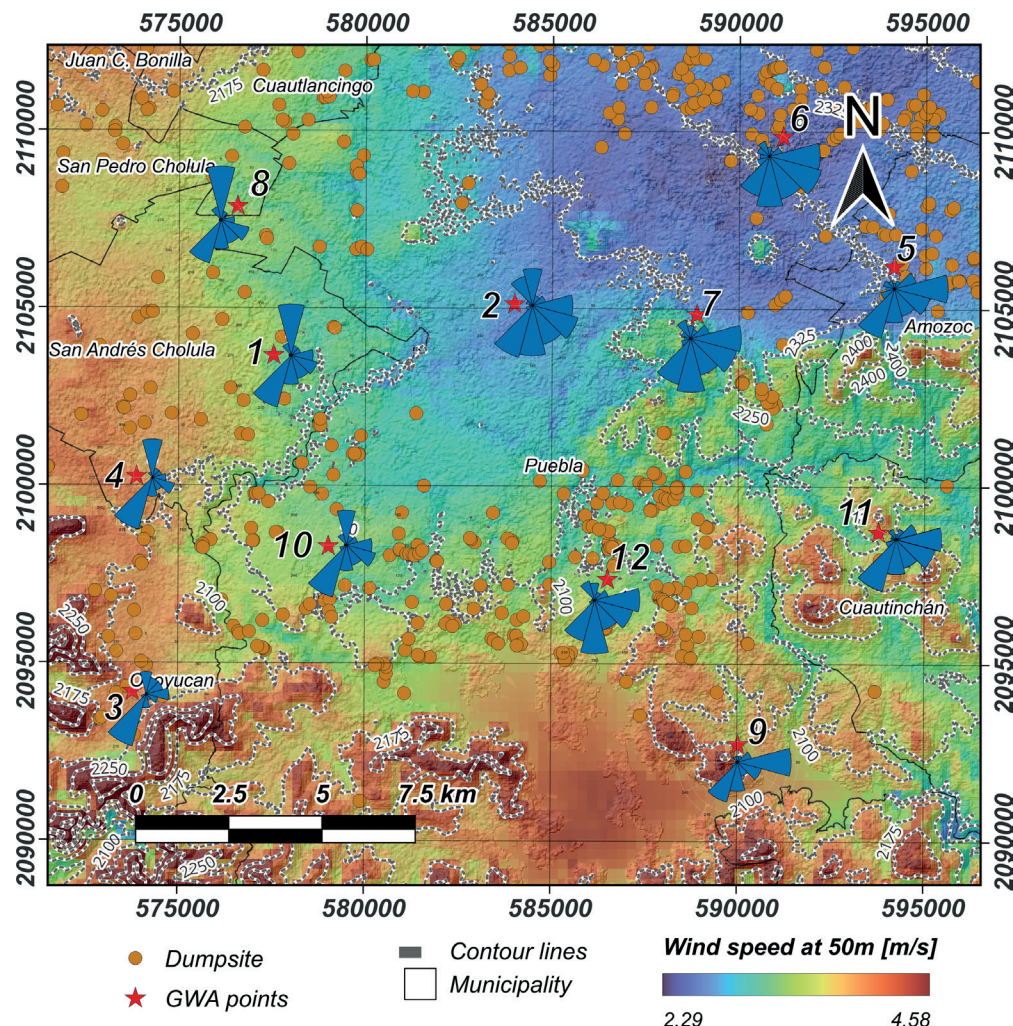


Figure 9. Map of the average wind speed in the Puebla-Tlaxcala metropolitan area. Average wind speed per site: 1) 3.53 m/s; 2) 2.83 m/s; 3) 5.19 m/s; 4) 4.81 m/s; 5) 3.34 m/s; 6) 3.14 m/s; 7) 3.13 m/s; 8) 4.37 m/s; 9) 4.78 m/s; 10) 3.45 m/s; 11) 4.47 m/s; 12) 4.31 m/s.

12 (Figures 9 and 10). The main wind directions in this area are toward the southwest and, to a lesser extent, toward the north and east (Figure 10).

The correlation between the lower speeds and methane accumulation in the northeast part of the city may be due to the topography of the area. Where the elevation of La Malinche volcano acts as a barrier, creating stagnant conditions that reduce the dispersion of polluting gases. In addition, lower speeds and the lack of a predominant flow direction may be causing gas recirculation in the northeast, due to the faster winds around it and its predisposition toward the southwest.

3.7. PCA

The principal component analysis (PCA) was applied to a dataset composed of seven variables: area, elevation, NDVI, LST, TVDI, CH₄, and distance to roadways for 2021 and 2022. In 2021, the first two principal components (PC1 and PC2) explained approximately 50% of the total variance (Table 4). When the third component (PC3) was included, the cumulative explained variance increased to 65% (Table 4). The correlations between the principal components and the original variables indicate the following patterns. The first component showed strong correlations with LST, TVDI, and distance to roadways (Table 4), indicating a relationship associated with water stress conditions. The second component was correlated with CH₄,

elevation, and TVDI (Table 4), suggesting a relationship between altitude and methane concentration. The third component was mainly associated with NDVI; however, a negative relationship was also observed within the first two components (Table 4), indicating an inverse relationship between LST and TVDI.

The PCA results for 2022 show that the first two principal components (PC1 and PC2) explained 58% of the total variance (Table 4). When the third component (PC3) was included, the cumulative explained variance increased to 73% (Table 4). The correlations between the principal components and the original variables were similar to those observed in 2021. The first component was mainly associated with LST, TVDI, and distance to roadways (Table 4). The second component showed a correlation between CH₄ and elevation (Table 4). The third component was primarily related to NDVI and the area of the solid waste sites (Table 4).

During the analysis, three main groups were identified with the following characteristics. In 2021, Group 1 (red) was characterized by high LST and TVDI values, as well as relatively high CH₄ concentrations and elevation (Figure 11A). Group 2 (green) showed high NDVI values associated with low LST and TVDI conditions (Figure 11A). Group 3 (blue) presented very low CH₄ concentrations and average elevation values (Figure 11A). In 2022, the previously identified groups maintained similar general characteristics. Group 1 (red) remained associated with high LST and TVDI values (Figure 11B). Group 2 (green)

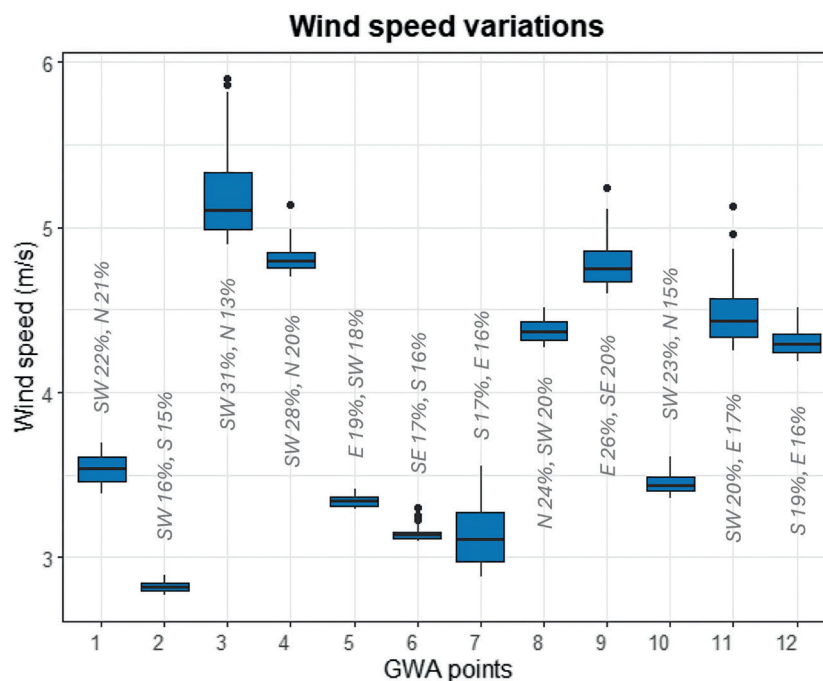


Figure 10. Wind speed and prevailing wind directions at sampling points in the study area. Prevailing directions and frequencies at recorded points: N – North; S – South; E – East; SW – Southwest; SE – Southeast.

Table 4. PCA results for 2021 and 2022 with three principal components.

2021				2022		
Variables	PC1	PC2	PC3	PC1	PC2	PC3
Area	0.48	-0.21	0.24	0.38	-0.09	0.59
Elevation	-0.24	0.63	0.17	-0.18	0.91	0.10
NDVI	-0.28	-0.13	0.92	-0.34	-0.06	0.70
LST	0.79	0.45	-0.09	0.92	0.26	-0.14
TVDI	0.58	0.62	0.27	0.91	0.22	-0.02
CH ₄	-0.40	0.71	-0.02	-0.16	0.91	0.14
Dist. Roads	0.61	-0.34	0.15	0.49	-0.21	0.41
Std. Deviation	1.36	1.29	1.02	1.49	1.35	1.02
Variance %	0.27	0.24	0.15	0.32	0.26	0.15
Cumulative variance	0.27	0.50	0.65	0.32	0.58	0.73

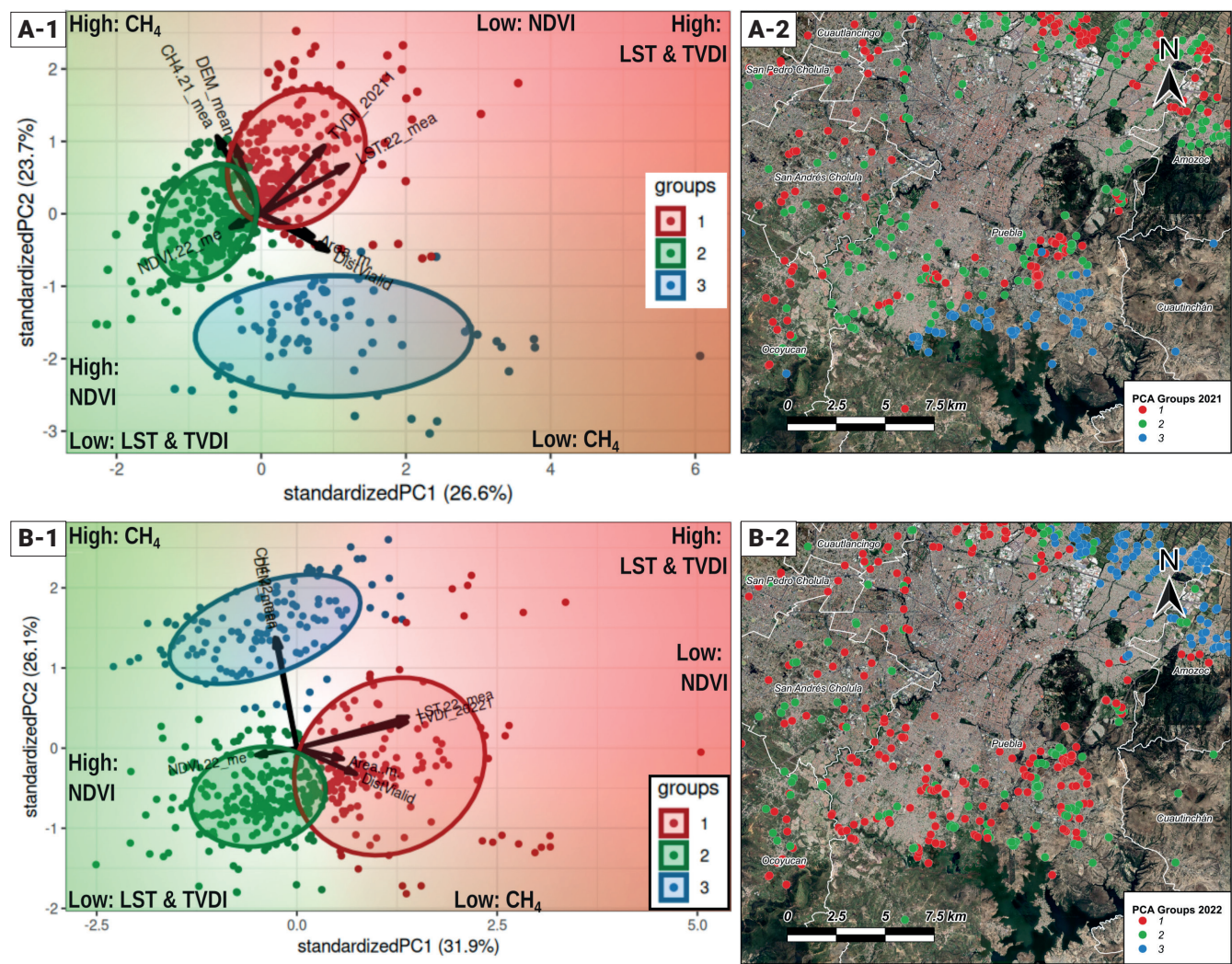


Figure 11. Principal Component Analysis of Environmental Variables Influencing Illegal Dumpsites SW Sites in Puebla. A) 2021, B) 2022.

continued to exhibit the highest NDVI values and lower water stress conditions (Figure 11B). In contrast, Group 3 (blue) showed a substantial change, exhibiting the highest methane concentrations and elevation values (Figure 11B).

3.8. Integrated Analysis of Methane and Environmental Variables

Atmospheric methane concentrations observed in Puebla City (mean values of approximately 1,884–1,900 ppb) are consistent with the urban background concentrations commonly reported for large metropolitan areas. Previous studies using TROPOMI data have shown that methane concentrations in urban environments are strongly influenced by local anthropogenic sources, atmospheric mixing conditions, and meteorological variability (Plant *et al.*, 2022).

The PCA results suggest that methane concentrations in the Puebla metropolitan area are not strongly controlled by local environmental conditions surrounding illegal dumpsites, such as vegetation cover or soil moisture. Groups characterized by contrasting NDVI, LST, and TVDI conditions showed similar methane concentrations in both analyzed years, indicating that the presence of vegetation or favorable moisture conditions for organic matter decomposition does not directly correspond to higher atmospheric methane levels. Furthermore, areas with the highest concentration of illegal dumpsites did not show a substantial increase in methane concentrations relative to the local background average.

Instead, methane variability appears to be more closely associated with topographic conditions. In 2021, higher methane concentrations were associated with elevated areas and moderate water stress, whereas in 2022, methane concentrations were concentrated within a cluster characterized by higher elevations. These results suggest that atmospheric circulation, topographic configuration, and local wind dynamics may influence methane transport and accumulation patterns in the study area. Previous studies using TROPOMI have demonstrated that methane concentrations detected in urban environments are strongly affected by atmospheric stability, boundary layer dynamics, wind circulation, and sensor detection thresholds, which can obscure direct relationships between small emission sources and detected atmospheric methane concentrations (Plant *et al.*, 2022; Jacob *et al.*, 2022).

The inverse relationship observed between NDVI and both LST and TVDI suggests that vegetation cover may partially mitigate thermal stress, whereas areas lacking vegetation are more susceptible to higher surface temperatures and lower soil moisture. Although methane concentrations did not show a strong relationship with vegetation conditions, NDVI values

revealed measurable differences among several illegal dumpsites, suggesting that these sites still produce localized environmental impacts. These impacts likely depend on factors such as waste deposition conditions, residence time, deposit age, and site persistence. It is possible that Group 1, characterized by lower NDVI values, corresponds to older dumpsites where vegetation degradation has occurred over longer periods. However, climatic variability may also influence vegetation conditions in the study area. February 2022 recorded higher precipitation (12.6 mm) than February 2021 (7.2 mm), which may have enhanced vegetation greenness independently of waste-related effects. Therefore, longer temporal analyses are necessary to better distinguish the environmental impacts associated with illegal dumpsites from broader climatic influences.

These results also highlight an important limitation of TROPOMI for detecting emissions from small and dispersed illegal dumpsites. Previous studies have shown that TROPOMI is more effective for identifying large and concentrated methane sources, such as sanitary landfills, industrial facilities, livestock emissions, and oil and gas infrastructure, than diffuse urban waste sites (Jacob *et al.*, 2022; Lauvaux *et al.*, 2022; Irakulis-Loitxate *et al.*, 2021). Additionally, although the area of each dumpsite was mapped, it was not possible to estimate waste volume, which may also influence methane production potential.

Finally, the PCA revealed a relationship between dumpsite area and distance to roadways, indicating that larger illegal dumpsites tend to be located farther from urban roads. This pattern suggests that areas with lower accessibility, reduced human presence, or limited surveillance may facilitate the accumulation and persistence of larger volumes of waste. Unlike large sanitary landfills, illegal dumpsites are generally smaller, spatially dispersed, and characterized by variable waste accumulation times and compositions, which can limit the detectability of methane anomalies at the spatial resolution of Sentinel-5P. Nevertheless, the environmental conditions associated with these sites, including elevated surface temperatures, reduced vegetation cover, and lower soil moisture conditions, may still indicate localized environmental degradation linked to waste accumulation processes (Richa-Singh, 2023).

4. Conclusions

This study successfully analyzed the interactions between environmental variables and illegal dumpsites in Puebla using advanced remote sensing techniques. The findings demonstrate that vegetation health, as indicated by NDVI, and surface temperature are distributed in ways controlled by the conditions at these illegal dumpsites. Notably, the analysis revealed that sites

with higher elevations tend to have greater methane concentrations, suggesting a direct correlation that warrants further investigation.

The Principal Component Analysis (PCA) provided valuable insights into the underlying patterns among various environmental factors. In 2021, two principal components captured 66% of the variance, primarily explaining correlations between methane levels, elevation, and distance to roads and vegetation health and land surface temperature. The PCA for 2022 indicated similar trends, with three components explaining 73% of the data variance, underscoring the persistent relationships among these variables over time.

From a geophysical standpoint, this study contributes to environmental geophysics by demonstrating how satellite-based measurements, such as land surface temperature, spectral vegetation indices, and atmospheric methane concentration, can be used to detect and characterize surface processes associated with anthropogenic waste accumulation. The integration of thermal infrared data (Landsat 7), high-resolution reflectance imagery (PlanetScope), and gas emissions monitoring (Sentinel-5P TROPOMI) supports the geoscientific assessment of energy exchange dynamics, surface degradation, and pollution dispersion in urbanized environments. This framework strengthens the application of geophysics to urban sustainability and risk management in Latin America.

Ultimately, this research contributes to a greater understanding of the complex dynamics in urban areas with poorly managed waste. The insights gained can guide future studies and inform stakeholders in developing sustainable dumpsites practices that mitigate environmental impacts and improve public health in Puebla City and other similar urban regions.

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Relationship of tropical cyclone wind-field and cloud-band metrics with extreme rainfall events over Mexico

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Abstract

Tropical cyclones (TCs) are the costliest hydrometeorological phenomena in Mexico, accounting for approximately 86.5% of the total disasters, where extreme rainfall is the main TC-associated hazard. This study analyzes two TC size metrics during the 2000–2024 period: R34, defined as the mean radius of 34-kt winds, and ROCLOUD, a satellite-derived metric that represents the spatial extent of TC cloud bands. These two TC size metrics are evaluated to determine their skill in detecting extreme precipitation events and identifying TC-associated disaster records from CENAPRED. We use diverse databases for this study: IBTrACS, MSWEP (used to define the 95th percentile of daily precipitation from May to November and identify extreme rainfall events), and official disaster declarations from CENAPRED. Municipalities are affected by TCs when they fall within the spatial domain of R34 (ROCLOUD). In addition, the SIAT-CT (Early Warning System for Tropical Cyclones by its acronym in Spanish) hazard index is computed during the study period to evaluate its performance in identifying extreme precipitation events induced by TCs. Our results show that at least 70% of TC-related disaster records are associated with tropical storms and tropical depressions. Extreme rainfall events frequently occur at distances greater than 300 km from the TC center, and ROCLOUD detects them better than R34. Radial curves indicate that ROCLOUD adequately covers the extreme events up to 500 km from the TC center, whereas the R34 skill to cover these events decays rapidly beyond 300 km. The SIAT-CT index, mainly based on R34, classifies the most extreme precipitation events into low-hazard categories, revealing important limitations for issuing TC rainfall-related warnings. Therefore, we conclude that SIAT-CT requires both TC wind-based metrics and TC cloud-band extension metrics in order to provide accurate warnings of TC-induced hazards and thus strengthen disaster risk reduction strategies prior to TC landfall.

Key words: Tropical cyclones, Extreme rainfall, Hazard assessment, Tropical Cyclone Size, Disaster management.

Resumen

Los ciclones tropicales (CTs) son los fenómenos hidrometeorológicos más costosos para México, ya que representan el 86.5% de los desastres, en los que la precipitación extrema es la principal amenaza hidrometeorológica. Este estudio analiza dos métricas del tamaño de CTs durante el periodo 2000-2024: el R34, que es el radio medio en el que se extiende la velocidad de 34 nudos y ROCLOUD, una métrica que mide la extensión de las bandas nubosas del CT. Ambas métricas se evalúan para determinar su capacidad de detectar la ocurrencia de eventos de lluvia extrema e identificar las declaratorias oficiales asociadas. Los datos utilizados son IBTrACS, precipitación de MSWEP, que sirvieron para obtener el percentil 95, y declaratorias de desastre de CENAPRED. Los municipios son afectados por CTs cuando están dentro del dominio espacial definido por R34 (ROCLOUD). También, se calculó el índice de peligro del SIAT-CT para evaluar su capacidad en identificar eventos extremos de precipitación inducidos por CTs. Los resultados muestran que el 70% de las declaratorias asociadas con CTs son debidas a tormentas y depresiones tropicales. Los eventos extremos de precipitación ocurren a más de 300 km del centro del CT, siendo ROCLOUD una mejor métrica para captar estos eventos en comparación con R34. Las curvas radiales confirman que ROCLOUD mantiene cobertura hasta 500 km desde el centro del CT, mientras que R34 decae rápidamente a partir de 300 km. El SIAT-CT, basado en R34, clasifica la mayoría de estos eventos extremos de precipitación en categorías de peligro bajas, lo que evidencia su limitación para emitir alertas relacionadas con la lluvia de los CTs. Por lo tanto, se concluye que el SIAT-CT requiere implementar tanto métricas de viento, como de la extensión de las bandas nubosas, para proporcionar alertamientos precisos de la amenaza que un CT induce y así fortalecer las medidas de prevención antes de que un CT toque tierra.

Palabras clave: Ciclones tropicales, Precipitación extrema, Evaluación de peligro, Tamaño de los ciclones, Manejo del desastre.

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1. Introduction

Hydrometeorological hazards (e.g., intense rainfall, floods, coastal storm surges, strong winds, droughts, and storms) account for approximately 46% of disasters recorded in Mexico during the 1900–2018 period (Alcántara-Ayala, 2019). Among these, Tropical Cyclones (TCs) are the most impactful hydrometeorological phenomena. TCs are responsible for 86.5% of the annual economic losses associated with disasters in Mexico, with extreme rainfall representing their most hazardous component (CENAPRED, 2019). In this context, extreme precipitation (or extreme rainfall) events are typically defined as accumulated daily rainfall exceeding the 95th or 99th percentile of rainy days over a baseline period of at least 20 years of daily data (Myhre *et al.*, 2019).

Although wind damage near the cyclone core can be severe, TC-related impacts often extend far beyond the storm center. In particular, intense rainfall in peripheral regions frequently triggers floods and landslides (Mandal *et al.*, 2020; Nolasco-Javier & Kumar, 2018). In recent years, multiple TCs made landfall or exerted significant influence over Mexican territory, highlighting the importance of studying key hazardous TC features, such as TC size, extreme rainfall, and associated flooding (Breña-Naranjo *et al.*, 2015; Dominguez *et al.*, 2021; Khouakhi *et al.*, 2017). In this sense, several studies divide the TC effects into two types: (i) direct effects within the TC center (e.g., extreme rainfall associated with the TC center and strong surface winds); and (ii) indirect effects that are far from the TC center, but are relevant for risk management actions (e.g. extreme rainfall located far from the TC center, but can cause floods and landslides). Hence, two complementary definitions of TC size have emerged: the TC inner size (confined to enclosing regions near the TC center) and the TC outer size (a region that extend far from the TC center, but the TC circulation can still be found) (Hidalgo *et al.*, 2020; Knaff *et al.*, 2014; Lu *et al.*, 2017; Pérez-Alarcón *et al.*, 2021).

Conventional approaches to characterizing direct TC effects have relied mainly on intensity-based indicators such as maximum sustained winds (MWS), as defined by the Saffir–Simpson scale (Kelman, 2013; Matyas, 2010). While these parameters provide an adequate metric for the destructive capacity of TC surface winds, they often fail to identify the broader and more complex hazard associated with rainfall extremes (i.e., indirect effects). Pérez-Estrada and Domínguez (2024) show that the spatial extent of TC rainbands, rather than peak TC intensity alone, plays a major role in modulating precipitation distribution and extreme rainfall events. This finding is consistent with previous studies highlighting the importance of rainband structure in shaping rainfall extremes (Domínguez & Magaña, 2018; Lavender & McBride, 2021; Morin *et al.*, 2024). Extreme precipitation events are usually the main natural causes of disasters. In Mexico,

complex topography raises rainfall-induced hazards and magnifies the resulting disaster impacts (see CENAPRED, 2019).

The operational Early Warning System for Tropical Cyclones (SIAT-CT, by its acronym in Spanish) in Mexico has four components: 1) scientific knowledge about TCs that affect Mexico (natural hazard component), 2) real-time monitoring of TCs (natural hazard component), 3) communication of public warnings that activate safety protocols (social component), and 4) local planning for well-organized social responses (social component) (CENAPRED, 2026). The SIAT-CT uses the radius of gale-force winds (R34) to determine the extent of damaging winds, which is considered a measure of the TC destructive potential (CENAPRED, 2019). R34 (which is a metric confined in the TC inner core) only provides a partial view of hazards that are confined near the TC center. This metric assumes that heavy precipitation occurs only near the TC center (Emanuel, 2005). Although this is partially true, R34 does not take into account this TC-induced hazard when computing its extent. Pérez-Estrada and Domínguez (2024) show that TC rainfall can extend hundreds of kilometers from the TC center, indicating that outer size can detect intense TC rainfall, which can be important for risk management. ROCLOUD was derived from satellite data (Global Precipitation Measurement; Skofronick-Jackson, 2020) and designed to identify convective structures far from the TC center that remain dynamically linked to the TC circulation.

Until now, a significant gap remains in combining wind metrics (such as R34), extreme precipitation events, coastal surges, and landslides with the SIAT-CT hazard level (Dominguez *et al.*, 2019). The creation of an integrated measure of TC hazardous impacts can benefit inhabitants by providing more accurate information for national and local risk management. The first step to achieving this integrated measure is to assess R34 skill in detecting extreme precipitation events and pointing out municipalities that are affected by TCs and therefore at risk of an emergency or disaster. Historical disaster declarations are obtained from the National Center for Disaster Risk Prevention (CENAPRED, by its acronym in Spanish). The SIAT-CT considers the value and spatial extent of R34 when assigning a TC hazard level. Specifically, the TC hazard-levels 4 and 5 require the most critical civil actions to be prepared for an emergency or a disaster. It is therefore important to evaluate whether the information previously provided by R34 corresponds to the disaster declarations issued by CENAPRED.

The main objective of this study is to assess the skill of R34 and ROCLOUD to detect the occurrence of extreme rainfall events and to identify disaster records in Mexico for municipalities that fall within their spatial domain. We use TC tracks, TC size metrics (i.e., R34 and ROCLOUD), a gridded precipitation dataset, and municipal-level disaster declarations from

CENAPRED during the 2000–2024 period. Our main scientific question is to determine the extent to which TC size metrics are associated with extreme rainfall events and local impacts for adequately representing TC hazards. The following section describes the datasets and methodology used for defining TC sizes, extreme rainfall, and hazard indices. Section 3 presents results about the analysis of TC size metrics with disaster records and evaluates the performance of the SIAT-CT hazard index in the context of extreme rainfall events. Section 4 mainly discusses the limitations of R34 and ROCLOUD. Finally, Section 5 summarizes the main conclusions and discusses the future directions for improving early warning systems in Mexico.

2. Data and Methods

2.1. TC databases: IBTrACS and ROCLOUD

The International Best Track Archive for Climate Stewardship (IBTrACS) integrates synoptic and operational information on TCs, ensuring consistency across basins and temporal continuity (Kenneth *et al.*, 2019; Knapp *et al.*, 2018). For this study, we selected TC tracks that occurred over the Eastern Pacific (EP) and North Atlantic (NA) basins during the 2000–2024 period.

Additionally, we used two TC size metrics: R34 and ROCLOUD. R34 is defined as the mean radius of gale-force winds (34 kt, $\sim 63 \text{ km h}^{-1}$) averaged across quadrants and expressed in kilometers. It has been used for operational purposes as a robust proxy for TC size and the spatial extent of hazardous winds within the TC inner core, and it is publicly available in the IBTrACS dataset (Bian *et al.*, 2021; Carrasco *et al.*, 2014; Lu *et al.*, 2017). The second metric is ROCLOUD, a recently developed satellite-based index that quantifies the outer radius of TC cloud bands (Perez-Estrada & Dominguez, 2024). The ROCLOUD algorithm combines TC wind fields with convective cluster locations. This technique uses GPM brightness temperature below 233 K (-40°C ; i.e., clouds at least 10 km above the surface) (Perez-Estrada & Dominguez, 2024). In this way, ROCLOUD tracks storm clouds that can produce extreme precipitation events (Cotton *et al.*, 2011). This brightness temperature is commonly used to define TC deep convection and intense rainfall (Zhou and Matyas, 2021). Unlike R34, which belongs to the inner structure, ROCLOUD determines the extent of outer convective bands where heavy rainfall may occur and where extreme precipitation events are most likely. In this way, ROCLOUD can provide more information on rainfall-induced hazards (e.g., floods and landslides) occurring far from the TC center. The dataset is publicly available at <https://data.mendeley.com/datasets/5bpzbwhynd/2>

2.2. Disaster declarations from CENAPRED

The disaster record database was compiled from CENAPRED from the period from 2000 to the present. It constitutes a repository of official municipal-level declarations of emergency, disaster, or climatological contingencies issued by the Mexican federal government in response to a wide range of natural and non-natural hazards (CENAPRED, 2019). This database represents the only official and legally binding record of disaster impacts in the country.

For this study, the declaration records were filtered to retain those explicitly and officially associated with TCs and their induced hazards. However, we found that some declarations from municipalities affected by TCs were not labeled properly, indicating an underestimation of TC-related declarations in CENAPRED records. So, we also included declarations from municipalities affected by TCs (although not labeled as TC-impacted in the CENAPRED database), whose locations fell within the TC spatial extent defined by TC size metrics (i.e., R34 or ROCLOUD) in this study. It is important to note that the number of TC-related declarations varies with the TC size metric. Records unrelated to hydrometeorological hazards (i.e., floods, landslides, extreme rainfall, and strong winds) were excluded.

2.3. Gridded Rainfall dataset

Historical rainfall was obtained from the Multi-Source Weighted-Ensemble Precipitation (MSWEP, v2.8) dataset, which merges in-situ gauges, satellite-based retrievals, and reanalysis products to provide spatially consistent estimates (Beck *et al.*, 2019). MSWEP offers a global record at 0.1° resolution and 3-hourly intervals, spanning 1979 to near-present. Its blended nature improves accuracy in regions with sparse gauge coverage, such as the Mexican Pacific and Atlantic coasts, which makes it particularly suitable for TC studies. Previous research has demonstrated that MSWEP captures localized convective rainfall structures associated with TCs (Morin *et al.*, 2024). We build on this research by using MSWEP as the primary dataset to characterize TC extreme rainfall events over Mexico during 2000–2024 period.

2.4. Method for comparing disasters records according to TC size metrics

The method workflow involves five stages: integrating rainfall data, TC size metrics, disaster records, hazard indices, and impact validation (Figure 1). The method is described as follows:

- 1. Detection of extreme rainfall.** Daily accumulated precipitation (rainy days) is computed from MSWEP 3-hourly data

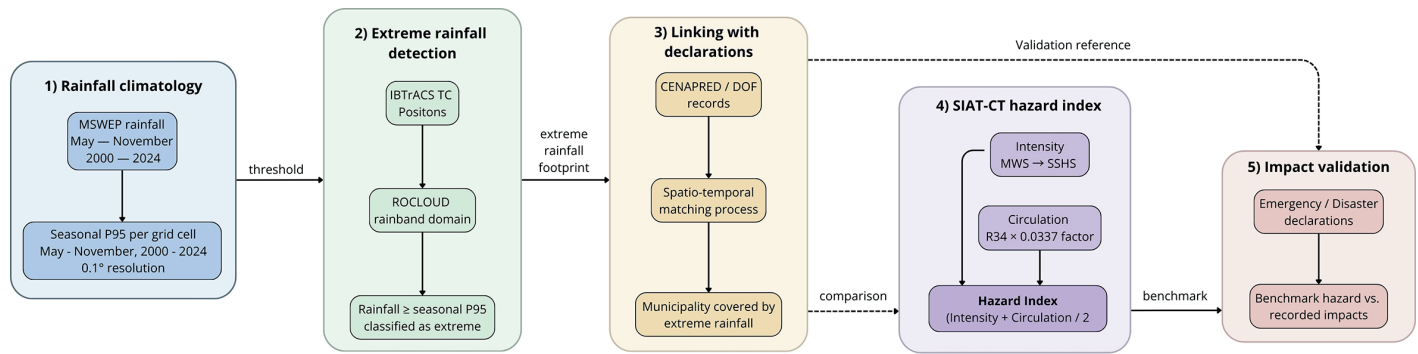


Figure 1. Method workflow. Steps include: (1) rainfall climatology from MSWEP database to define local P95 thresholds; (2) detection of extremes within the ROCLOUD domain; (3) spatio-temporal linking with CENAPRED disaster declarations; (4) calculation of the SIAT-CT hazard index from intensity and circulation; and (5) validation of hazard indicators against official impact records. MWS and SSHS mean Maximum Wind Speed and Saffir-Simpson Hurricane Scale, respectively.

during the 2000–2024 period. Only daily precipitation that occurs from May 1st to November 30th (the months when TCs are active) is considered for this study (hereafter, seasonal precipitation). Following Dominguez *et al.* (2021), we estimate a unique climatological threshold, defined as the 95th percentile of all daily rainy days for each 0.1° MSWEP grid cell during the 2000–2024 period, using only seasonal precipitation. This threshold varies regionally across Mexico and poses a hazard to local infrastructure, rivers, and dams. In this way, extreme precipitation events are defined as days with precipitation at or above the 95th percentile and represent a hazard. Their quantity provides a better estimation of TC-impacts rather than only considering 3-hourly rainfall (Dominguez *et al.*, 2021). It is important to note that this percentile does not account for any maximum monthly or annual rainfall amount in its calculation.

2. Identification of TC-induced rainfall events. For each 6-hourly TC position reported by IBTrACS, the R34 and ROCLOUD metrics (TC size metrics) are used to define the TC radial domain of influence. TCs usually move and change their latitude and longitude every 6 hours, but TC positions and size metrics are not averaged daily, as their radial domain of influence can affect different grid cells on the same day. If TC positions remain nearly unchanged for 24 hours, this does not affect the identification of extreme precipitation events, as locations remain invariant. TC-induced extreme rainfall events are attributed to each TC position. These events are flagged when precipitation within this domain exceeds the local 95th percentile. This procedure generated an event-based catalog of extreme rainfall associated with TC positions, thereby linking TC structure with hydrometeorological extremes.

- 3. Linking TC disaster declarations with extreme rainfall events.** To establish a connection between extreme rainfall events and disaster records, a spatiotemporal matching routine was applied. First, only CENAPRED disaster declarations that explicitly identified TCs as the triggering cause were retained. These TC-related declarations were used to define the set of affected municipalities and candidate impact locations. Extreme rainfall events were then identified when rainfall equaled or exceeded the climatological 95th percentile threshold during the TC lifetime. For each TC-related declaration, extreme rainfall points occurring within the corresponding municipality and temporally associated with the TC were retained as TC-related extreme rainfall impacts.
- 4. SIAT-CT hazard index. The SIAT-CT was historically reproduced using IBTrACS data.** The SIAT-CT defines hazard levels by combining: (a) intensity, derived from the MWS according to the Saffir–Simpson Hurricane Scale (SSHS), and (b) circulation, which is computed as the average of four R34 radii (i.e., northeast, northwest, southwest, and southeast), multiplied by 0.0337, with a maximum of five (CENAPRED, 2019). The hazard index is the mean of both scales, yielding a color-coded warning ranging from minimal to maximum hazard (i.e. from 0 to 5).
- 5. Impact validation.** Municipal-level declarations of emergency and disaster were considered as the impacts. These declarations provided an operational benchmark against which the physical hazard descriptors (ROCLOUD, R34, extreme rainfall, and SIAT-CT) could be evaluated. This comparison allowed assessing not only the ability of size and rainfall metrics to explain hazard, but also their consistency with institutional disaster recognition in Mexico.

3. Results

3.1. Official disaster declarations linked to TCs

Official disaster records in Mexico provide unique information about the social impacts of hydrometeorological hazards. During the 2001–2024 period, TCs accounted for 52% of all official hydrometeorological declarations. In other words, declarations that are only officially labeled as TC-associated by CENAPRED are presented in this section. Emergencies were the dominant category (56%), followed by disasters, while climatic contingencies represented only a minor share (approximately 3%; Figure 2a). Although other hazards, such as floods, droughts, and cold fronts, also triggered numerous declarations, the relative weight of TCs underscores their central role in shaping Mexico's disasters. Out of 34,453 accumulated declarations associated with all types of phenomena during our study period, 90.3% corresponded to hydrometeorological hazards, of which at least 10,543 (30.6%) were directly or indirectly linked to TCs. It is worth noting that the disaster declaration database does not, in many cases, explicitly specify the phenomenon linked to the declaration. The latter leads to an underestimation of the true contribution of TCs.

The geographical distribution of official declarations records reveals regional contrasts (Figure 2b). The state with the most declarations is Baja California Sur. Along the southern Pacific and Gulf of Mexico coasts, municipalities issued more than 30 declarations during the study period. These regions are frequently affected by TCs that made landfall mainly as tropical storms or tropical depressions. The high number of declarations indicates that these regions have a combination of high exposure with high social vulnerability, which enlarges impacts (Dominguez *et al.*, 2021). On the other hand, in sparsely populated regions, such as northern states, the number of declarations is markedly lower. This result is partly due to a minor hazard incident, but it could also point to institutional bias or underreporting, as prior evaluations of the Mexican disaster management system have suggested.

Declarations related to tropical storms and tropical depressions are more common than those for hurricanes (i.e., categories 1 and 2) and major hurricanes (i.e., categories 3, 4, and 5), with tropical depressions and tropical storms jointly accounting for approximately 70–83% of all TC-associated declarations, depending on the basin that the TC originates, during the study period (Figure 3a). Despite their weak surface winds, these systems are highly effective at producing widespread rainfall and flooding

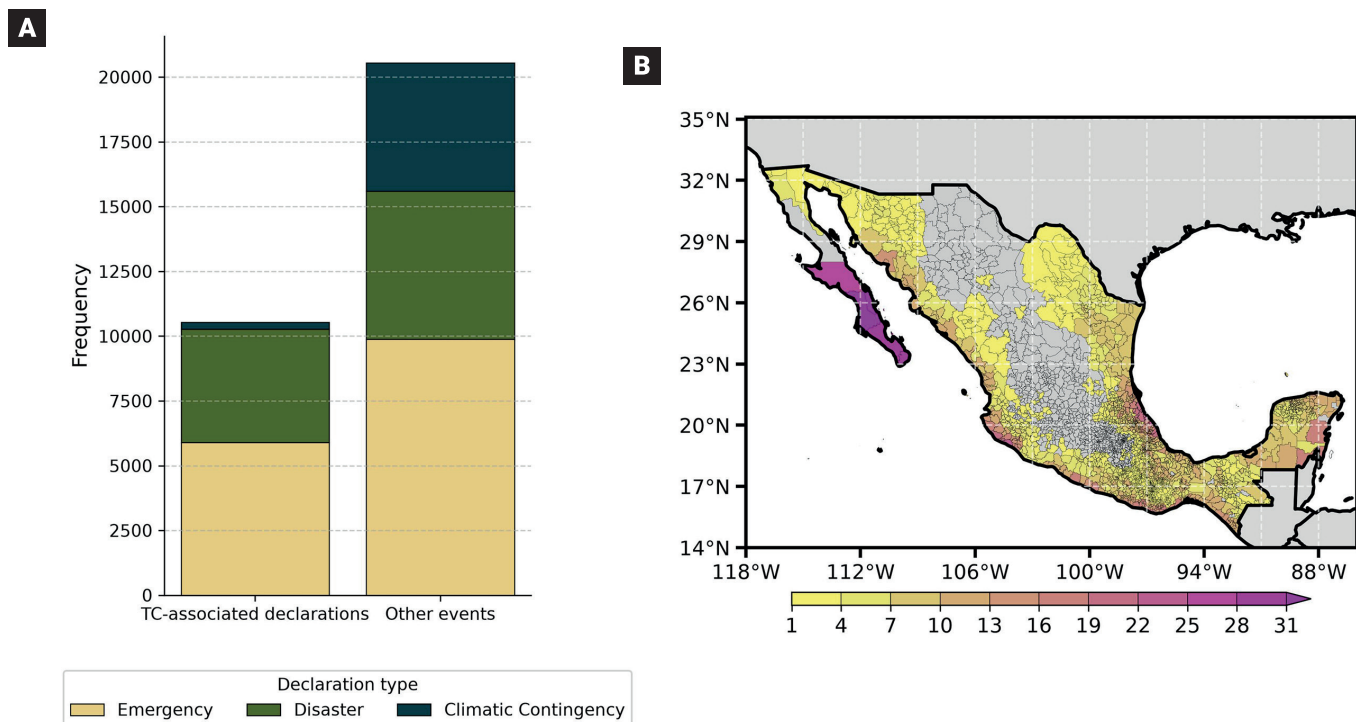


Figure 2. Disaster declarations from the National Center of Disaster Prevention (CENAPRED in Spanish) over Mexico during the 2001–2024 period. (A) Distribution of emergency, disaster, and climatic contingency declarations, separating those associated with TCs from other hydrometeorological hazards. (B) Spatial distribution of the total number of official declarations associated with TCs at the municipality level. Shaded areas represent the number of declarations.

while interacting with complex topography (Breña-Naranjo *et al.*, 2015; Dominguez *et al.*, 2022; Hidalgo *et al.*, 2020; Touma *et al.*, 2019). The latter results in a high number of emergency and disaster declarations, mainly over coastal municipalities adjacent to mountain ranges, such as the central-southwestern coasts of Mexico (Figure 2b). By contrast, major hurricanes account for less than 10% of all TC-driven declarations, which is associated with their lower frequency of landfall in Mexico (Dominguez & Magaña, 2018). It points out that weak TCs (e.g., tropical depressions and tropical storms) represent a frequently severe hazard to Mexico, although they belong to the lower categories of the Saffir-Simpson scale.

Apart from the Saffir-Simpson scale, the outer TC size can also be relevant for determining the rainfall-induced hazards. However, these outer sizes vary spatially and are not fully correlated with the Saffir-Simpson scale (i.e., its size metric, R34). To illustrate this point, we explore the values of R34 and ROCLOUD during Hurricane Alex in 2010 at 00:00 UTC on 1 July 2010 (Figure 3b). The gale-force wind radius, represented by R34, adequately captures the TC inner core. However, the four ROCLOUD quadrants (i.e., northeast, northwest, southwest, and southeast), which are based on the extent of TC circulation and its cloud bands (Perez-Estrada & Dominguez, 2024), are not close to the TC center, indicating a wider domain of TC affect-

tation. This example underscores the limitations of wind-only metrics (as in R34) in representing all TC-affected areas and, consequently, hydrometeorological hazards associated with TC passages. In this sense, ROCLOUD is a more realistic proxy for rainfall-induced impacts. This result suggests that R34 could underestimate TC indirect effects (i.e., flooding and landslides), which arise from outer TC convective bands that extend beyond the TC inner wind field. This finding is consistent with the results reported by Hidalgo *et al.* (2020).

3.2. Relationship of TC sizes with extreme precipitation events

The 95th percentile of daily precipitation from May 1st to November 30th was computed as a first step to obtain the mean annual extreme rainfall produced by TCs (Figure 4a). Mean annual extreme rainfall associated with TC passages varies spatially across Mexico. It is interesting to mention that mean annual extreme rainfall is constrained to the location of mountain ranges, where landslides commonly occur (Breña-Naranjo *et al.*, 2015; Dominguez & Magaña, 2018). For example, mean annual extreme precipitation produced by TCs can reach up to 100 mm/day, mainly southwards 24°N over mountain ranges. Central Mexico that is located inland is not affected by TC extreme precipitation.

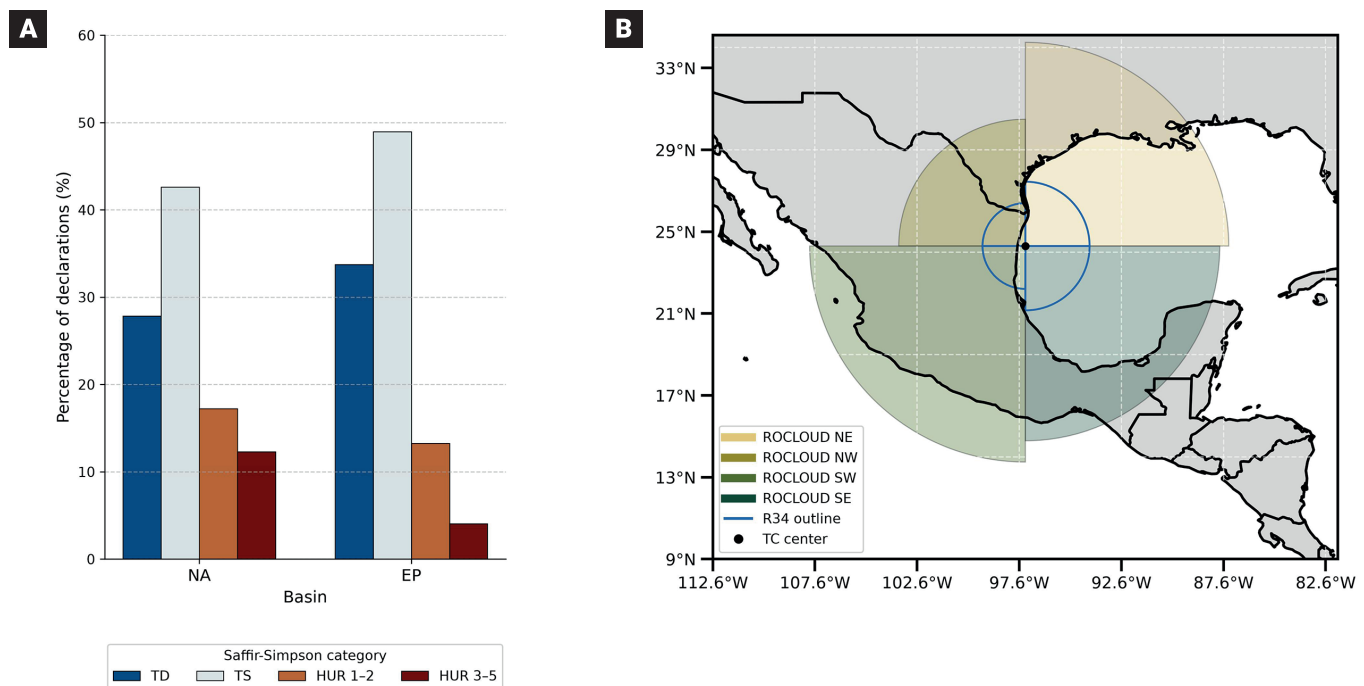


Figure 3. (A) Saffir-Simpson scale (TD: tropical depression, TS: tropical storms, HUR 1-2: hurricanes category 1 and 2, HUR 3-5: hurricanes category 3, 4 and 5) and its percentage of official TC-associated declarations according to the North Atlantic Ocean (NA) and Eastern Pacific Ocean (EP), (B) R34 and four quadrants of ROCLOUD radii (NE: northeast, NW: northwest, SW: southwest, SE: southeast) during Hurricane Alex in 2010 at 00:00 UTC on 1 July 2010. A black dot represents the TC center.

We estimate the mean distance from TC centers at which extreme rainfall events are produced by TC circulation (Figure 4b). In general, extreme rainfall events over the Gulf of Mexico and Caribbean Sea coast are induced by TCs whose centers are at least 200 km from landmass. It is worth noting that these events can be induced by TC centers located 350–500 km from landmass along the Mexican Pacific coast, mainly in Oaxaca, Michoacan, Colima, Jalisco, and Sinaloa states (Figure 4b). These

results show that extreme rainfall events are not confined to the TC inner core but are present in regions that are distant from the TC center, which suggests that ROCLOUD provides broader coverage of extreme precipitation than R34 (see extension of R34 in Figure 3b).

We compute the percentage of extreme precipitation events identified by ROCLOUD (R34) (Figure 5) as a division between the number of TC extreme precipitation events identified by RO-

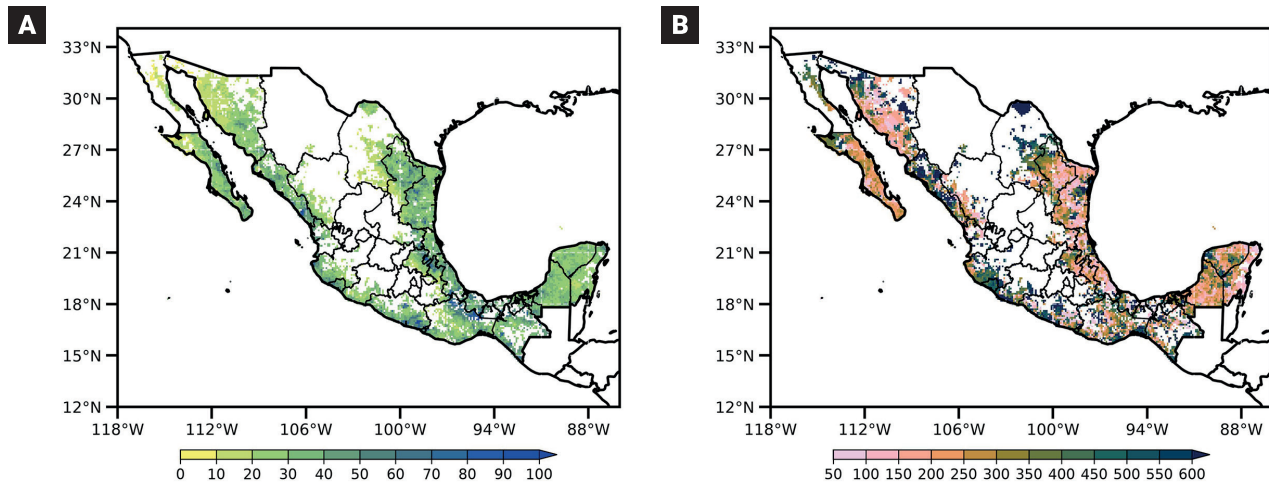


Figure 4. (A) Mean annual extreme precipitation (mm/day) associated with TCs in Mexico during the 2000–2024 period. Extreme precipitation is defined as daily rainfall equal to or exceeding the climatological 95th percentile, which is based on rainy days from May 1st to November 30th during the study period. (B) Mean radial distance (km) from the TC center to the location of extreme precipitation.

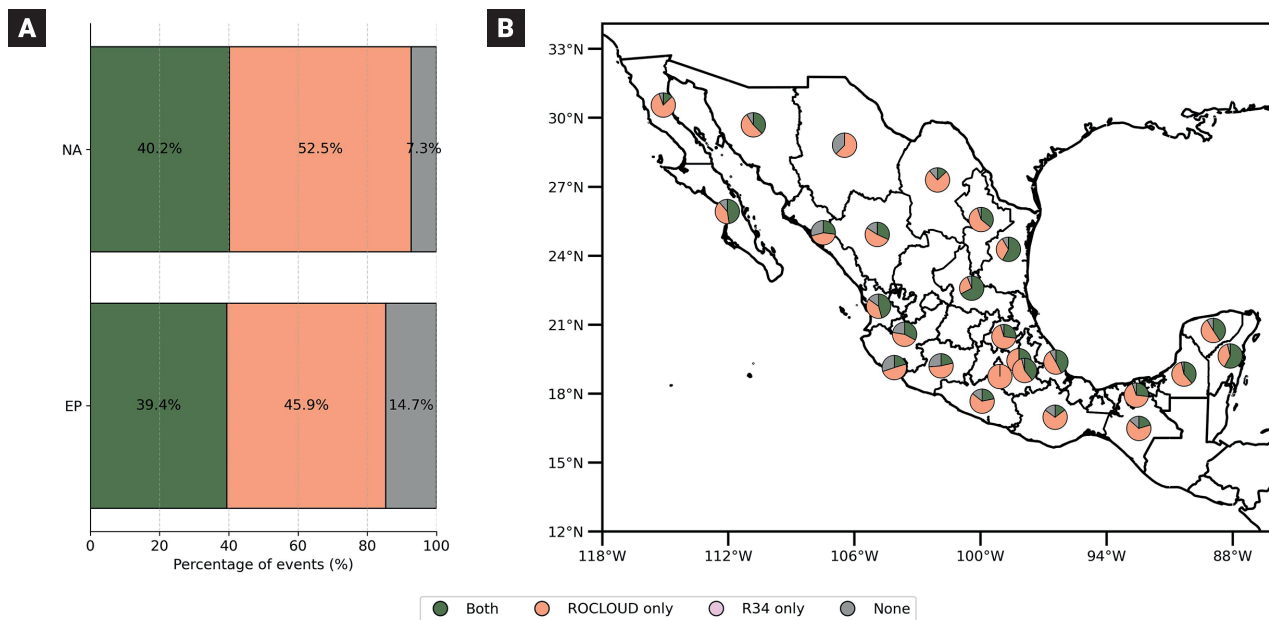


Figure 5. (A) Percentage (%) of extreme rainfall events per basin and (B) percentage (%) at the state level of extreme rainfall events captured by ROCLOUD, R34, both, or none TC size metrics during the 2000–2024 period.

CLOUD (R34) and the total number of TC extreme precipitation events, as shown in Figure 4a. ROCLOUD by itself captures between 52.5% and 45.9% of extreme precipitation events induced by TCs that originate over the NA and EP, respectively (Figure 5a). What is interesting to mention is that around 40% of all extreme rainfall events over both basins (at a national level) are captured by ROCLOUD and R34. By contrast, R34 only captures less than 1% of all events on both basins at a state level. This indicates that ROCLOUD encompasses nearly all events identified by R34, while additionally capturing a substantial number of extreme

rainfall events not detected by R34. It is important to bring out that 7-14% of all events are not detected by any TC size metrics. When analyzing the percentage of both metrics at the state level, most regional TC extreme events are identified by ROCLOUD or by its combination with R34. This result underlines that R34 lacks the skill to include hazards induced by TC extreme rainfall (e.g., landslides, floods, and extreme precipitation that extends beyond the TC inner core) at a state level.

We also analyzed the percentage of precipitation extreme events per grid (0.1° of spatial resolution) covered by ROCLOUD

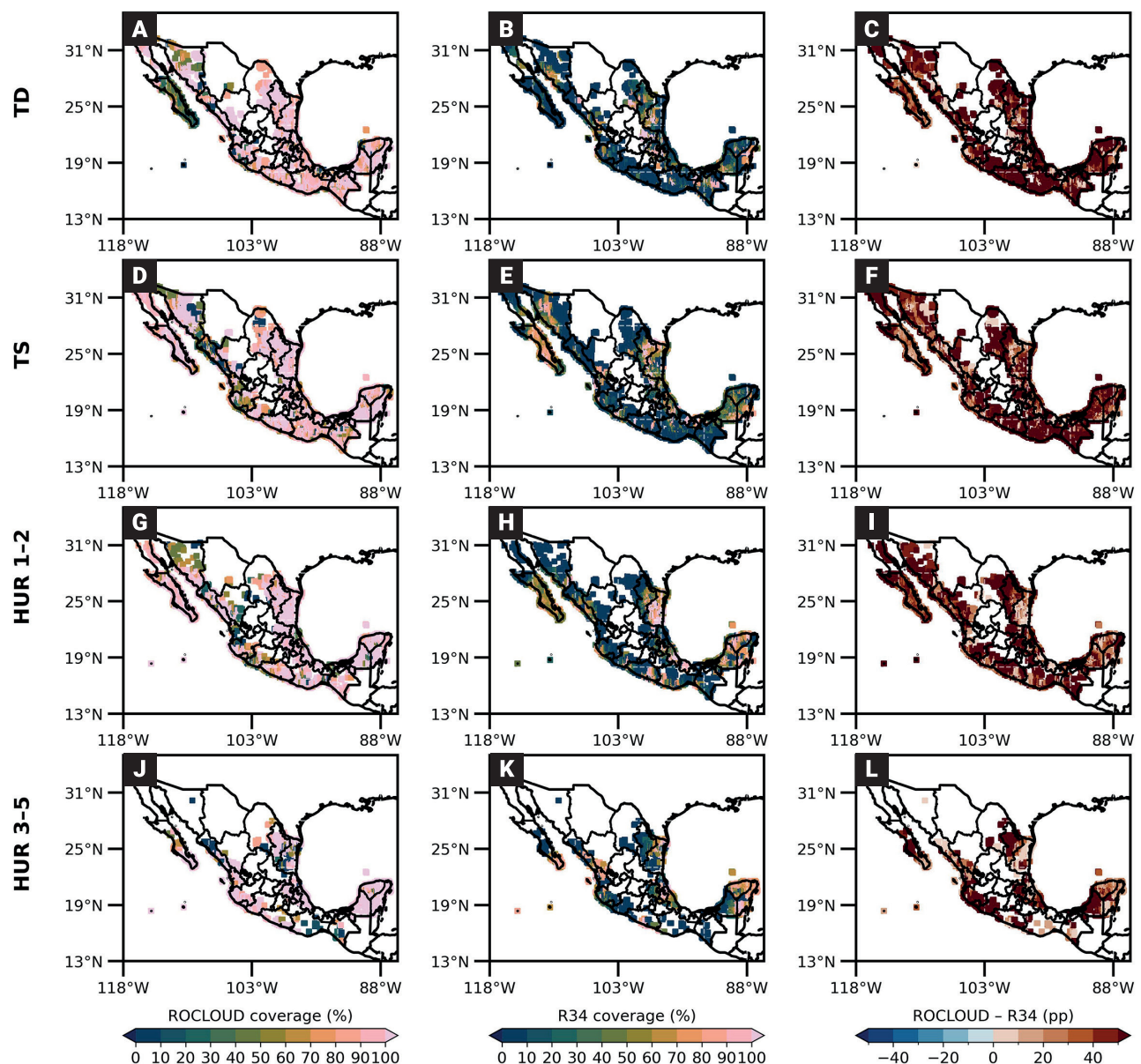


Figure 6. Percentage (%) of TC extreme precipitation events covered by ROCLOUD and R34. The left column represents the percentage of events covered by ROCLOUD, the middle column represents the percentage covered by R34, and the right column represents their difference (ROCLOUD – R34 in percentage), according to the Saffir-Simpson scale intensity during the 2000-2024 period. TD: tropical depressions, TS: tropical storms, HUR 1-2: hurricanes category 1 and 2, HUR 3-5: hurricanes category 3 and 5.

and R34, according to the Saffir-Simpson scale intensity during the 2000–2024 period (Figure 6). These results demonstrate that ROCLOUD identifies more than 80% of all extreme events near the Gulf of Mexico and Caribbean Region in all TC intensities. However, the percentage decreases near the Mexican Pacific coast, mainly in the tropical depressions and tropical storms that affect the Baja California Peninsula and Sinaloa. This behavior is likely linked to small TC radii affecting this region and their typically offshore trajectories of tropical depressions over cold EP waters. This small TC radius affects a minor fraction of the Mexican landmass, consistent with previous TC size statistics (Perez-Estrada & Dominguez, 2024). By contrast, R34 covers only around 40% of all extremes across coastal and inland regions, revealing that this TC wind-based metric does not present the same spatial pattern as the TC extreme rainfall distribution. R34 adequately captures the extreme events in more than 60% for specific locations and intensities. For instance, ~60% of extreme precipitation events induced by tropical storms are covered by R34 in Baja California Sur. In addition, ~65% of extreme precipitation events induced by hurricanes (category 1 and 2) and major hurricanes (category 3, 4, and 5) are captured by R34 over the Yucatan Peninsula, Tamaulipas, and the south of Baja California Sur, regions that are commonly affected by intense TCs (Dominguez Sarmiento, 2022). It is worth noting that states located inland (e.g., Aguascalientes, Zacatecas, Guanajuato, and EdoMex) are not affected by TC extreme rainfall events, as shown in Figures 4 and 6.

While making differences between ROCLOUD and R34 percentages of coverage, the differences show that ROCLOUD

detects more than 30%–40% of extreme rainfall events when compared to R34, depending on the region and TC intensity. The lowest difference is found in the major hurricanes, where percentages can be up to 20%. This result suggests that R34 properly detects extreme precipitation induced by major hurricanes rather than by weak TCs (e.g., tropical depressions, tropical storms). In this sense, ROCLOUD performance is relatively consistent across TC intensity categories, although slight reductions in coverage are observed for extreme precipitation induced by tropical depressions over Baja California Sur, likely reflecting regional storm characteristics rather than methodological limitations.

The conditional probability of coverage is estimated for ROCLOUD and R34 using diverse radial distances from the TC center and Saffir-Simpson scale intensities (Figure 7). For quantifying uncertainty, a nonparametric bootstrap resampling was applied at the 95% confidence level for all TC intensities. Confidence bands are narrower at short distances for all TC intensities, using both R34 and ROCLOUD, due to the sample size. However, these bands widen beyond ~700 km, where fewer extreme precipitation events were observed, mainly for ROCLOUD. These results point out important different features between the two TC size metrics. The curves for R34 nearly drop rapidly and approach zero within 300–350 km from the TC center, showing no significant differences in TC intensity or basin. On the other hand, ROCLOUD covers around 90–100% of all extreme rainfall events at 550–650 km from the TC center, depending on the basin. TC intensity bands are noticeably different beyond 700 km in both basins. Hurricanes and major hurricanes show slightly higher persistence than tropical depressions and storms,

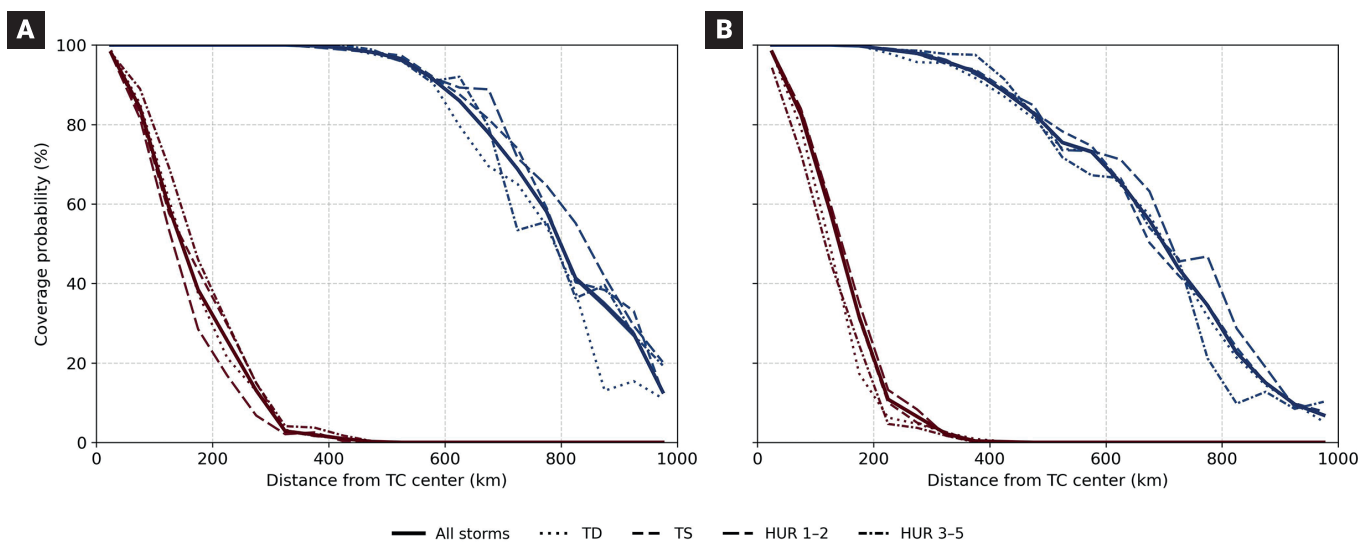


Figure 7. Radial coverage probability of extreme precipitation events by distance from the TC center. Curves show empirical coverage for ROCLOUD (in blue) and R34 (in brown) for TCs that originate in (A) the NA and (B) the EP, with bootstrap confidence intervals and separation by Saffir-Simpson scale intensity during the 2000–2024 period.

particularly over EP. However, the 95% bootstrap confidence bands widen in this radial range, indicating that such differences are not statistically distinguishable from the aggregated curve.

Figures 6 and 7 reveal that ROCLOUD consistently detects a more accurate extreme rainfall distribution induced by all TC intensities than R34. Therefore, the TC size metric R34 systematically underestimates rainfall-induced hazards in regions located more than 300 km from the TC center, and remains strongly dependent on TC intensity. These findings suggest that the operational SIAT-CT, by relying exclusively on a TC wind-based metric, systematically underrepresents rainfall-induced hazards. Consequently, this structural limitation constrains early warnings for rainfall-related impacts, particularly in regions of complex topography such as the Mexican mountain ranges. Incorporating TC cloud-based size metrics into SIAT-CT would substantially improve the robustness and reliability of its early warning protocols.

3.3. Assessment of the SIAT-CT hazard index

The SIAT-CT requires both surface wind intensity and TC circulation (defined by R34) to compute its hazard levels from 0 (no hazard) to 5 (maximum hazard) (i.e., SIAT-CT hazard index). This index is operationally relevant for Mexico, thus its performance in representing precipitation-driven impacts requires careful assessment. The relationship between SIAT-CT hazard

level and TC extreme rainfall distribution shows no considerable differences (Figure 8a). Median values of extreme rainfall events (i.e., around 30 mm/day) remain invariant from hazard levels 0 to 5, with a large overlap in the interquartile ranges. This finding shows that extreme rainfall events can occur at any hazard level, undermining the SIAT-CT skill to properly warn rainfall-induced hazards. What is more, Figure 8b demonstrates that ~70% of extreme rainfall events fall within TC-hazard levels 1 and 2, with very few events indicated in TC-hazard levels 4 and 5, levels that indicate that actions to face emergency and disaster should be prepared. This bias suggests that the SIAT-CT systematically underestimates the rainfall-induced hazard associated with TC passages, and it can assign a minor hazard level to TCs that can be extremely dangerous due to their extreme precipitation.

Considering all historical SIAT-CT hazard-level warnings during the 2000-2024 period, we find that the mean SIAT-CT hazard level, as defined by R34, ranges from 0 to 2 in most of the country (Figure 9a), even in regions with recurrent landslides (e.g., Michoacan, Guerrero, Oaxaca, Chiapas, Veracruz, and Puebla) (Domínguez-Morales *et al.*, 2016; Mendoza-López *et al.*, 2002). It is remarkably important to mention that each hazard level corresponds to safety protocols that society implements before TCs impact Mexico. So, low-level hazards indicate low impact and suggest that emergency or disaster declarations will not occur (Perez Estrada *et al.*, 2024). In addition, the percentage of extreme rainfall events classified as low-level hazards

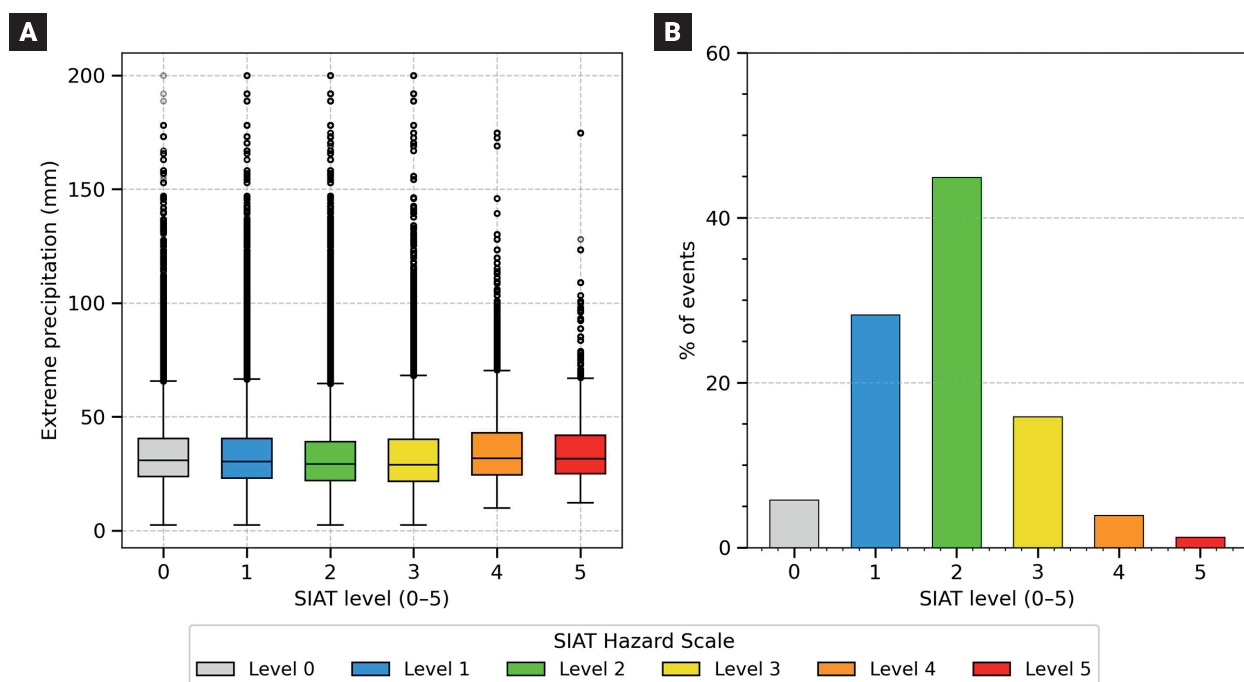


Figure 8. Distribution of TC extreme precipitation events by SIAT-CT hazard levels. (A) Boxplots of TC extreme precipitation (mm/day) across hazard level 0-5. (B) Relative frequency (%) of extreme precipitation events assigned to each hazard level during the 2000-2024 period.

(i.e., from 0 to 2) varies from 50 to 80% in most of the country (Figure 9b), revealing a consistent spatial underestimation of the TC rainfall-induced hazard. This serious underestimation is particularly evident over mountain regions, where orographic TC convection is enhanced.

These findings exhibit that the SIAT-CT hazard index based on R34 is a valuable tool for warning hazards associated with winds from TCs whose Saffir-Simpson intensity is hurricane categories 3, 4, or 5. However, this index critically fails to warn hazards associated with weak TCs (i.e., tropical depressions and tropical storms), despite at least 70% of TC-related disaster declarations being associated with these hydrometeorological phenomena (Figure 3a).

By contrast, ROCLOUD can identify extreme rainfall events produced by TCs from both basins, demonstrating to be a better metric that considers TC circulation and cloud bands in its computing (Perez-Estrada and Dominguez, 2024), when compared to R34. Incorporating wind and rainfall metrics, such as ROCLOUD, into the operational SIAT-CT framework would enhance the system's capacity to adequately assign hazard levels and properly implement safety protocols in advance of TC landfall. The latter will help reduce TC-driven disasters and accurately communicate TC hazards.

4. Discussion

R34 is used for describing the radial extent of gale-force winds near the TC inner core and the main metric for issuing the TC-hazard level for the SIAT-CT. However, this radial extent does not encompass most TC rainfall-induced hazards.

This limitation is particularly relevant for Mexico, where tropical depressions and tropical storms are the most common TCs affecting the country (Figure 3a). Their TC-related impacts are associated with extreme rainfall, flooding, and landslides rather than destructive winds alone. Extreme rainfall events produced by TCs frequently occur beyond the inner-core wind field at distances greater than 300 km from the TC center (far from distances defined by R34, Figure 7). This mismatch between wind radii and extreme precipitation events limits SIAT-CT's operational capability to warn of TC-level hazards, especially in coastal and mountainous regions exposed to outer rainbands.

ROCLOUD is a metric that represents the outer cloud-band structure with rainfall-producing convection. We demonstrated that this TC metric detects 52.5% and 45.9% of extreme precipitation events induced by TCs originating over the NA and EP, respectively (Figure 5a), compared with less than 5% detected by R34. ROCLOUD is used here as a physical proxy for the spatial domain where extreme precipitation events induced by TC circulation can occur. One limitation of the ROCLOUD algorithm is that it considers only convection with brightness temperatures below 233 K. Although most extreme precipitation can occur in the cloud-cover area detected by ROCLOUD, some extreme precipitation events produced by clouds lower than 10 km above the surface (e.g., cumulus congestus and nimbostratus) might be neglected. These lower clouds can only produce extreme precipitation events in the presence of local cloud microphysical processes (Khadke *et al.*, 2024), which are outside the scope of this study. Furthermore, ROCLOUD is affected by satellite cloud-top detection, temporal sampling, quadrant-based radial simplification, and displacement between cold cloud shields and surface precipitation (Sun *et al.* 2017).

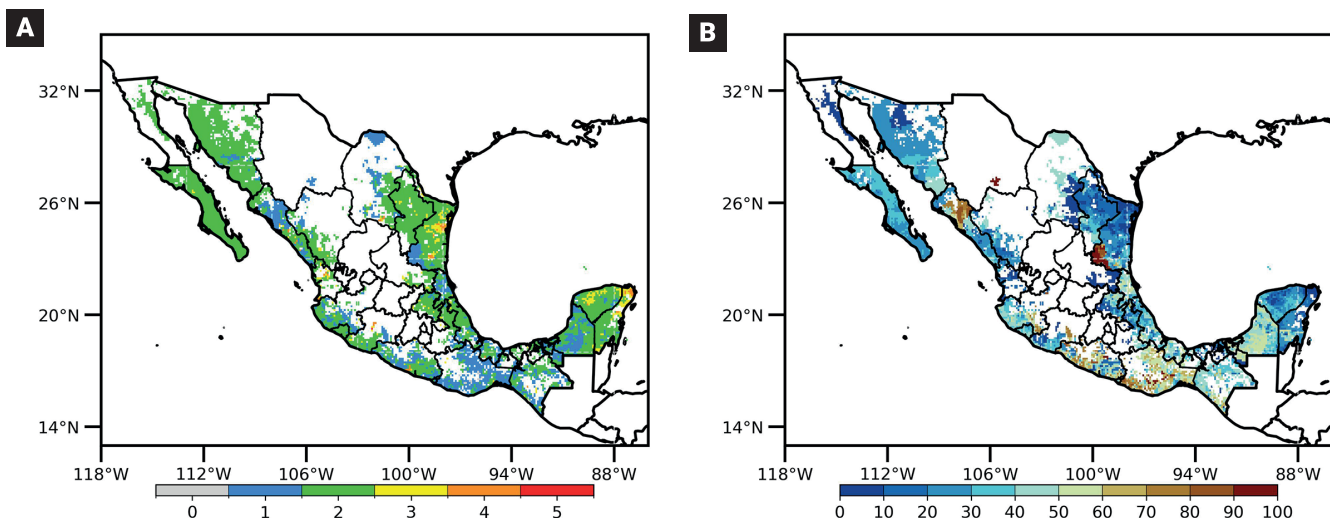


Figure 9. (A) Historical mean of SIAT-CT hazard level (B) percentage (%) of extreme rainfall events assigned to SIAT-CT low-level hazards (0–2) during the 2000–2024 period.

Declarations from CENAPRED (i.e., emergency and disaster declarations) are institutional governmental records, so their issuance is influenced by reporting capacity, administrative criteria, emergency management practices, and governmental decisions regarding financial support (CENAPRED, 2022). The latter shows uncertainty about the annual number of official TC-associated declarations, which is likely underestimated by the Mexican government. Our results show that no official TC-associated declarations are reported inside the country (Figure 2b), although TC extreme precipitation occurs (Figure 4a), demonstrating this underestimation. SIAT-CT commonly assigns TC-hazard levels 1 and 2 for TC extreme precipitation (Figure 8b). These levels indicate a minimum and low hazard, respectively, and do not correspond to actions to be taken in response to an emergency or disaster (CENAPRED, 2019). The SIAT-CT has been operational since 2000, and the formula for computing TC-level hazards has not been modified since then. Here, we show that TC-level hazards associated with extreme precipitation events have been underestimated by SIAT-CT (e.g. Tropical Storm Cristobal 2020 as shown in Perez-Estrada *et al.*, 2024). Thus, the current SIAT-CT wind-centered formula is insufficient to characterize the full spatial extent of rainfall-related impacts.

Our findings demonstrate that ROCLOUD can be used as an important input variable for classifying TC hazard levels. The future approach should consider forecasting TC size (i.e., ROCLOUD), TC wind speed, and daily accumulated precipitation to incorporate this information into SIAT-CT. This study can also be useful to other countries affected by TCs, whose topography is complex.

5. Conclusions

This study demonstrates that TC wind-based metrics, such as R34, partially include TC rainfall-induced hazards, which are responsible for at least 70% of the TC-associated disaster declarations in Mexico. The latter represents a critical limitation for the performance of SIAT-CT, which assigns low-level hazards to TC extreme rainfall events. Consequently, the implementation of proper safety protocols is constrained by a bias in SIAT-CT. On the other hand, TC size metrics that consider both TC circulation and cloud band extension, such as ROCLOUD, are robust proxies for capturing the extreme rainfall events occurring beyond 300 km from the TC center. The use of this TC size metric could help warn of flooding and landslides, hazards arising from TC extreme rainfall rather than TC surface winds. These findings underline the need to improve the SIAT-CT by incorporating TC precipitation-driven metrics alongside traditional wind-based metrics into its operational framework.

A future modern SIAT-CT should integrate multiple components of the TC structure (i.e., inner-core wind fields and outer-size rainfall) to properly warn direct and indirect hazards to society. The latter will enhance alignment between TC hazard levels and disaster declarations, an advance urgently needed, and help Mexican media properly communicate the TC hazard level. Accurate TC early warnings can provide a reliable basis for local decision-makers to reduce disaster risk, particularly in municipalities with high-vulnerability populations (Dominguez *et al.*, 2021).

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The Silverman Minimum in the mid-18th century: a better assessment

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Abstract

Understanding historical solar activity is essential due to its impacts on Earth's climate, space weather, stellar behavior, and human health. The recent centuries have witnessed two pronounced solar minima through telescopic observations of sunspots: the Maunder Minimum in the 17th century and the Dalton Minimum in the 19th century. Proxies for solar activity like aurora records and cosmogenic isotopes reveal other solar minima in earlier centuries. In addition, Silverman (1992) found a previously unrecognized minimum around 1765 using mid-latitude aurora observations. The present work studies the annual number of days when auroras were recorded in mid-latitudes (AD), the official sunspot number series (SN) and sunspot numbers derived from cosmogenic radioisotopes (SN_{14C}) to evaluate if the Silverman Minimum can be considered a notable minimum in solar activity beyond auroral records. Our results reveal a pronounced minimum in AD during 1755–1766 and in SN_{14C} during 1747–1752, whereas just a non-pronounced minimum in SN over 1745–1761. While the uncertainty of the datasets could explain part of the differences in the results, comparisons with other periods of reduced solar activity suggest that the differences among series also occur. Moreover, non-pronounced SN minima can be compatible with pronounced minima in other solar activity indices, such as AD, without implying a mischaracterization of the datasets.

Key words: Sunspot number, Aurora, Cosmogenic isotopes, Solar cycle, Historical records.

Resumen

Comprender la actividad solar histórica es esencial debido a sus impactos en el clima de la Tierra, el clima espacial, el comportamiento estelar y la salud humana. En los siglos recientes se han registrados dos mínimos solares de actividad solar muy reducida a través de observaciones telescópicas de manchas solares: el Mínimo de Maunder en el siglo XVII y el Mínimo de Dalton en el siglo XIX. Los indicadores de la actividad solar, como los registros de auroras y los isótopos cosmogénicos, revelan otros mínimos solares en siglos anteriores. Además, Silverman (1992) encontró en observaciones de auroras en latitudes medias un mínimo previamente no reconocido alrededor de 1765. El presente trabajo estudia el número anual de días en que se registraron auroras en latitudes medias (AD), la serie oficial del número de manchas solares (SN) y la serie del número de manchas solares derivada de radioisótopos cosmogénicos (SN_{14C}) para evaluar si el Mínimo de Silverman, además de ser un mínimo en los registros de auroras, puede considerarse un mínimo destacado de actividad solar. Nuestros resultados revelan un mínimo pronunciado en AD y SN_{14C} durante el período 1755–1766 y 1747–1752, respectivamente, mientras que solo un mínimo no pronunciado en el SN durante el período 1745–1761. Si bien la incertidumbre de las variables podría explicar parte de las diferencias en los resultados, los análisis y las comparaciones con otros períodos sugieren que las diferencias entre variables también ocurren en otros mínimos y que los mínimos no pronunciados en el SN pueden ser compatibles con mínimos pronunciados en otras variables relacionadas con la actividad solar, como AD, sin implicar una caracterización errónea de ninguna de las variables.

Palabras clave: Número de manchas solares, Aurora, Isótopos cosmogénicos, Ciclo solar, Registros históricos.

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1. Introduction

Understanding past solar activity is crucial for multiple areas of research. It provides insights on the long-term behavior of the Sun and its influence on Earth's climate (Haigh, 2007). Knowledge of past solar activity is also essential for space weather research (Hanslmeier, 2007). Solar storms can disrupt communication systems, satellites, and power grids, and studying historical solar activity helps develop better forecasting models and mitigation strategies. Understanding past solar activity contributes to our knowledge of long-term solar behavior and stellar evolution, as the Sun is a typical star (Brun *et al.*, 2015). In addition, solar activity modulates the Earth's magnetic field and cosmic ray fluxes with implications for technological systems and potential links to human health (Mishev and Usoskin, 2018).

The behavior of different manifestations of solar activity, such as sunspots, and proxies as auroras and cosmogenic isotopes, provide a better knowledge of the long-term variability (Usoskin, 2023). Regarding direct solar records, sunspot counting with telescope has been available only since 1610 (Arlt and Vaquero, 2020; Clette *et al.*, 2023) although some observations of sunspots have been made with naked eye in the last two millennia (Wittmann and Xu, 1987; Yau and Stephenson, 1988). Proxies of solar activity such as aurora records are also available for two millennia (Fritz, 1873) and reconstructed series using cosmogenic radioisotopes covers even more than ten millennia (Solanki *et al.*, 2004).

Telescopic observations of sunspots indicate the occurrence of two notable minima in the last four centuries: the Maunder Minimum, which was a grand minimum period (Eddy, 1976; Usoskin *et al.*, 2015; Vaquero *et al.*, 2015; Carrasco *et al.*, 2021), spanning the second half of the 17th century and the beginning of the 18th century, and the Dalton Minimum in the early 19th century, which was a prolonged period of low solar activity but not considered as a grand minimum (Hayakawa *et al.* 2020; Carrasco, 2022). Moreover, using cosmogenic isotopes as a proxy, other grand solar activity minima were identified in the last millennium (Usoskin, 2023): Oort (around the year 1000), Wolf (around 1300), and Spörer (around 1500) Minimum.

Silverman (1992) studied the secular variation of northern aurora observations for the period 1450-1948. In the series of annual number of days on which auroras were observed in mid-latitudes (south of 55°N in geographic latitude), he detected the known Spörer, Maunder, Dalton, and 1901-1913 prolonged solar activity minima. In addition, he found an evident previously unrecognized minimum around 1765 (Silverman Minimum, hereafter). Subsequent works such as Vázquez, Vaquero, and Gallego (2014) and Ptitsyna, Demina, and Tyasto (2018) confirmed the existence of the Silverman Minimum in auroral records. The

above raises an important question to understand the long-term solar activity behavior. Is the Silverman Minimum a minimum only in auroral records? Apparently, according to the sunspot number (Vaquero, 2007; Clette *et al.*, 2014; Muñoz-Jaramillo and Vaquero, 2019), there is no deep minimum in the middle of the 18th century. On the other hand, high-resolution series (annual cadence) of cosmogenic radionuclides currently available (Usoskin *et al.*, 2021) can provide another point of view about the Silverman Minimum.

The objective of this work is to analyze if the Silverman Minimum could be considered a solar activity minimum or only a minimum in auroral records. For this purpose, an aurora data series is studied as well as two sunspot number series: the official sunspot number series and the sunspot number series reconstructed from cosmogenic radioisotopes measurements. These data series used in the present work are described in detail in Section 2. The methodology followed to process the data is explained in Section 3. Section 4 contains the discussion of our results and, lastly, the main conclusions of our work are presented in Section 5.

2. Data

Our work is based on the analysis of three data series characterizing solar activity and covering, at least, the 18th century, in which the Silverman Minimum was found. Firstly, northern aurora observations in mid-latitudes (south of 55°N in geographic latitude) are considered since that minimum was detected in that type of observations, which are directly correlated with solar activity (Silverman, 1992). For this purpose, we have used the catalog of northern aurora observations in mid-latitudes by Křivský and Pejml (1988) supplemented by Křivský (1996), which covers the period 1000-1900. The data of this catalog were digitized, processed and quality controlled by the National Oceanic and Atmospheric Administration (NOAA) and National Aeronautics and Space Administration (NASA). The catalog consists of a list of dates (year, month and day arranged in chronological order) when auroras have been observed, together with the source of the information, obtained from sets of observations and all the catalogs (or their corrections) published to its date taking into account criteria to avoid overestimating the number of aurorae. From it, we have computed the annual number of days when northern auroras were recorded in mid-latitudes (AD), i.e., at geographic latitudes below 55°N. Secondly, we have used the official annual sunspot number series (SN) (version 2, see Clette and Lefèvre, 2016) provided by the Sunspot Index and Long-term Solar Observations (SILSO, <https://www.sidc.be/SILSO>), which is the longest time series based on direct observations reflecting

long-term solar activity. The annual sunspot number series covers the period from 1700 to the present. Thirdly, we have chosen the high-resolution series by Usoskin *et al.* (2021) of annual sunspot number derived from ^{14}C concentration measurements in tree rings ($\text{SN}_{14\text{C}}$), which spans the period 971–1899, since this series achieves an annual resolution. Table 1 lists the annual values of the above three series for the 18th century, whereas the whole dataset used in the present study is available at the request of those interested.

3. Results

To assess the solar activity level during the 18th century, including the Silverman Minimum around the middle of that century, the AD, SN and $\text{SN}_{14\text{C}}$ data series for that century are presented in Figure 1 (upper panel). In addition, 11-year running averages of the three series are shown in Figure 1 (lower panel) to show their long-term behaviors removing the influence of the 11-year solar cycle. For a broader overview, Figure 2 includes the above data series for a longer time interval, namely the period 1600–1950. This allows comparison of the Silverman minimum with known solar activity minima and thus evaluates whether it can be considered a notable minimum.

4. Discussion

4.1. The Silverman Minimum

For the 18th century, the AD series presents a temporal evolution similar to those of SN and $\text{SN}_{14\text{C}}$ series (Figure 1, upper panel). In the case of annual series (Figure 1, top panel), rises and falls of the indices approximately follow the 11-year solar activity cycle with cycles of higher maximum amplitudes in the second half of the century with respect to the first one. The correlation coefficient between the AD and SN series is $r = 0.60$ and it is $r = 0.63$ between AD and $\text{SN}_{14\text{C}}$, both with a p -value < 0.001 .

Regarding the 11-year smoothing series (Figure 1, bottom panel), the three indices present a higher solar activity level in the second half of the century compared to the first one. The average of the AD series for the first half of the century is 15.6, whereas it is 32.7 in the second half. In the case of SN, it is 65.4 in the first half and 91.0 in the second half, whereas it is 35.8 versus 78.8, respectively, for $\text{SN}_{14\text{C}}$. Thus, while the average for the second half of the century with respect to the first one increased higher than 100 % in the case of the AD and $\text{SN}_{14\text{C}}$ series, it is 39 % in the case of the SN.

Besides the Maunder Minimum, which ended in the early 18th century, and the Dalton Minimum, which started near the end of that century, the AD series shows a remarkable minimum occurring around mid-18th century (Figure 1, bottom panel). This minimum was first identified by Silverman (1992). The smoothed AD series decreased from values around 20–30 near 1730 to approximately 5 around 1760. A minimum is also evident in the smoothed $\text{SN}_{14\text{C}}$ series in the mid-18th century, where values declined from approximately 50–80 during 1720–1740 to values near 0 around 1750. In contrast, the decrease in the smoothed SN series is less pronounced since values decreased from approximately 80–90 around 1730 to about 65–70 around 1750.

The minimum periods in $\text{SN}_{14\text{C}}$ and AD present differences in the year of their lowest value (1749 for $\text{SN}_{14\text{C}}$ and 1762 for AD) and they do not overlap with each other (1747–1752 for $\text{SN}_{14\text{C}}$ and 1755–1766 for AD). Minimum periods were defined as intervals around local minima during which the smoothed series remained below a threshold value given by: $X_{\min} + 0.25(X_{\max} - X_{\min})$, where $X_{\min}$ is the value of the local minimum around mid-18th century and $X_{\max}$ is the maximum value of the period before the local minimum. On the other hand, the year of the minimum in the SN series (1746) is closer than that occurred in the $\text{SN}_{14\text{C}}$ series than in the AD series. Also, all the years of the local minimum for the SN overlap with years of the minimum for the SN series (1745–1761), whereas only some years of the AD overlap with those for the SN series. Figure 1 (bottom panel) provides a clear view of these differences between the three solar activity indices studied.

The temporal offset between the minima in the AD and $\text{SN}_{14\text{C}}$ series, as well as the longer duration of the minimum in AD, may reflect the different physical nature of the two proxies. While $\text{SN}_{14\text{C}}$ mainly reflects long-term modulation of cosmic rays by the solar magnetic field, AD is associated with geomagnetic disturbances capable of producing visible mid-latitude auroras and is therefore more sensitive to the occurrence of intense oronal mass ejections (CME) and strong high-speed solar wind streams (HSS). Consequently, relatively small differences in solar magnetic complexity and geomagnetic effectiveness during weak solar cycles may produce larger and more persistent variations in AD than in $\text{SN}_{14\text{C}}$, contributing to the temporal displacement and broader minimum observed in the AD series during the SM.

4.2. Comparison with Other Periods of Reduced Solar Activity

A comparison of the Silverman minimum with others observed in the telescopic era can be seen in Figure 2. The similarities among the indices remain when the time interval

Table 1. List of annual number of days when northern auroras were observed at geographic latitudes below 55°N according to Křivský and Pejml (1988) supplemented by Křivský (1996) [AD], SILSO annual values of the official sunspot number (version 2) [SN], and annual sunspot number obtained from ^{14}C concentrations in tree rings provided by Usoskin *et al.* (2021) [$\text{SN}_{14\text{C}}$] for the 18th century.

YEAR	AD	SN	$\text{SN}_{14\text{C}}$	YEAR	AD	SN	$\text{SN}_{14\text{C}}$
1700	0	8.3	19.8	1751	12	79.5	19
1701	0	18.3	1.2	1752	12	79.7	9
1702	3	26.7	2.8	1753	5	51.2	-17.6
1703	0	38.3	-2.2	1754	6	20.3	-31.3
1704	7	60	-2.8	1755	2	16	-18.7
1705	2	96.7	8.4	1756	5	17	20.5
1706	0	48.3	2.5	1757	2	54	70.8
1707	15	33.3	0.9	1758	3	79.3	101.4
1708	5	16.7	-1.2	1759	8	90	138.8
1709	2	13.3	-9	1760	2	104.8	134.3
1710	1	5	-8.3	1761	3	143.2	90.6
1711	1	0	-9.5	1762	2	102	66.8
1712	0	0	-6.3	1763	2	75.2	61.7
1713	3	3.3	4.5	1764	7	60.7	68.5
1714	0	18.3	9.7	1765	1	34.8	88.9
1715	2	45	13.3	1766	0	19	94.8
1716	17	78.3	2.1	1767	1	63	83.5
1717	13	105	-3.5	1768	8	116.3	62
1718	18	100	12	1769	24	176.8	20.5
1719	30	65	12.3	1770	23	168	-4.4
1720	17	46.7	12.9	1771	17	136	-15.5
1721	23	43.3	34.5	1772	20	110.8	-8.6
1722	44	36.7	63	1773	39	58	21.1
1723	18	18.3	75.6	1774	40	51	65
1724	6	35	73.1	1775	20	11.7	126.6
1725	16	66.7	84.8	1776	32	33	189.3
1726	18	130	98.4	1777	69	154.2	224.8
1727	18	203.3	106	1778	74	257.3	237.6
1728	44	171.7	119.4	1779	112	209.8	207.9
1729	20	121.7	103.3	1780	41	141.3	159.9
1730	50	78.3	81	1781	69	113.5	132.2
1731	18	58.3	51.5	1782	51	64.2	123
1732	40	18.3	0.9	1783	76	38	112.7
1733	15	8.3	-35	1784	49	17	126.1
1734	24	26.7	-40.1	1785	41	40.2	156.5
1735	20	56.7	-4.2	1786	104	138.2	182
1736	37	116.7	38.1	1787	128	220	193.2
1737	38	135	80.6	1788	127	218.2	191.7
1738	14	185	151.1	1789	101	196.8	180.6
1739	28	168.3	190.2	1790	76	149.8	127.5
1740	5	121.7	174.6	1791	75	111	73.8
1741	35	66.7	137.9	1792	55	100	47.5
1742	16	33.3	97.1	1793	18	78.2	8.2
1743	15	26.7	51.9	1794	8	68.3	-20.6
1744	5	8.3	6.4	1795	6	35.5	-29.5
1745	3	18.3	-14	1796	3	26.7	-29.9
1746	18	36.7	-14.7	1797	14	10.7	-16.4
1747	25	66.7	-6.8	1798	1	6.8	-7.7
1748	10	100	13	1799	5	11.3	20.8
1749	17	134.8	38.8	1800	7	24.2	54.3
1750	41	139	29.7				

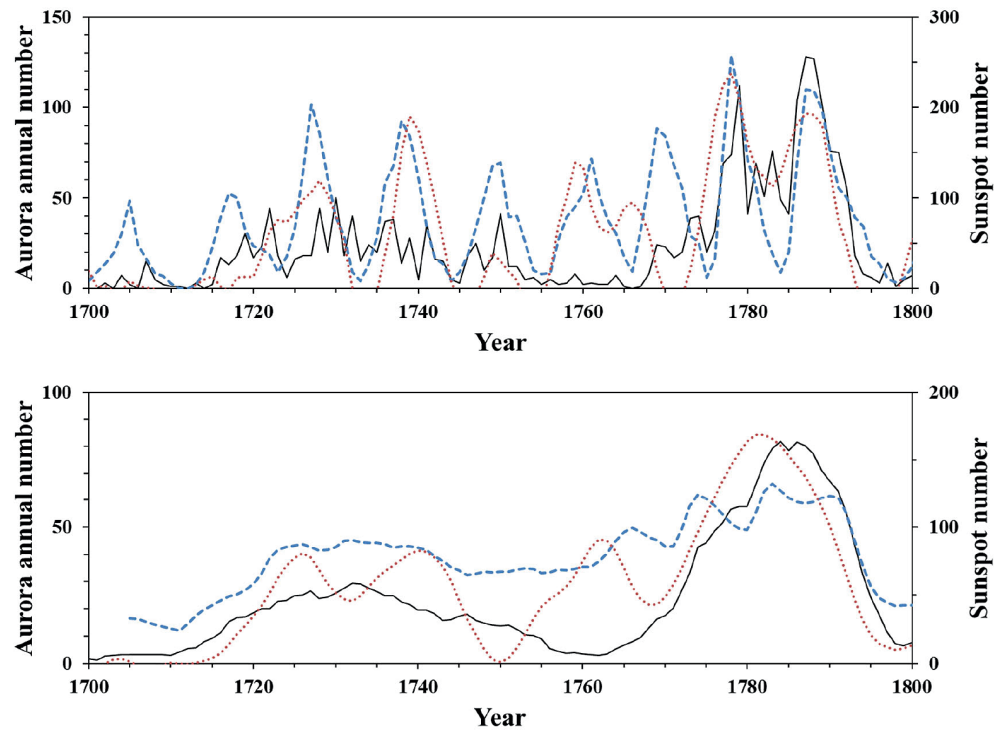


Figure 1. (Upper panel) Annual number of days when northern auroras were sighted at geographic latitudes below 55°N (AD) according to Křivský and Pejml (1988) supplemented by Křivský (1996) in black continuous line, annual values of the sunspot number (version 2) by SILSO (SN) in blue dashed line, and annual sunspot number derived from ^{14}C concentrations in tree rings ($\text{SN}_{14\text{C}}$) by Usoskin *et al.* (2021) in red dotted line for the 18th century. (Lower panel) The same three series after computing 11-year running averages.

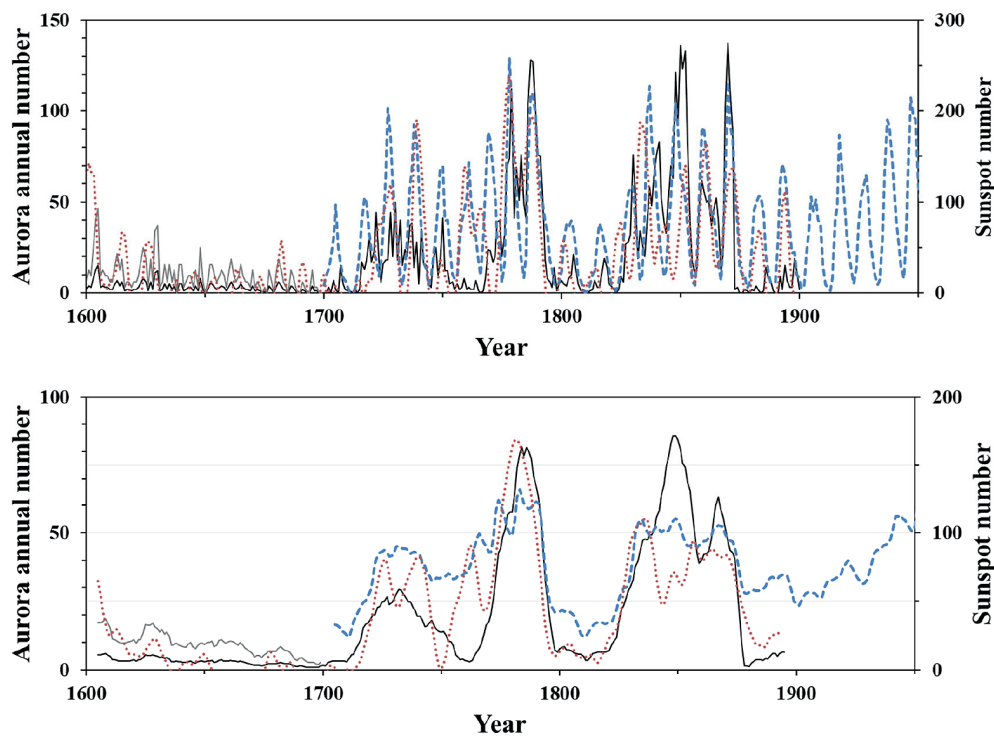


Figure 2. Same as Figure 1, but for the period 1600-1950. The different temporal coverage of the three series reflects their data availability: AD is available from 1600 to 1950, SN from 1700 onward, and $\text{SN}_{14\text{C}}$ over a reconstructed interval dependent on cosmogenic data. (Upper panel) Annual values. (Lower panel) 11-year running means.

is extended from the 18th to the 19th century (Figure 2). Regarding the annual series, the three studied indices exhibit approximately 11-year cycles (Figure 2, top panel). Moreover, regarding annual data of the 18th and 19th centuries, the correlation coefficient between AD and SN_{14C} is $r = 0.61$ (which is similar to that found only for the 18th century), although $r = 0.52$ for AD-SN (which is lower than 0.60, that is the value obtained using data 18th century data). The p-value is lower than 0.001 in both cases.

A similar behavior in the solar activity level of the three indices can be seen regarding 11-year smoothing (Figure 2, bottom panel). Regarding AD series, the depth of the Silverman Minimum (SM) is similar to that corresponding with the Maunder Minimum (MM), the Dalton Minimum (DM) and the Gleissberg minimum between the late 19th century and the early 20th century (GM). In particular, the AD values during those minima reached values as low as in the range of 1–5 in all cases. The difference with the other minima is in the time extension during those values were observed. For example, regarding the SM, values below 10 (threshold used for AD values in Section 4.1) were observed for the period 1755–1767, whereas they were below 10 during all the MM (except a few years at the onset with values around 10), for 1798–1822 in the DM and from 1877 to the end of the studied AD (1894) for the GM.

In relation to the SN_{14C} series, one can see that the mid-18th century minimum has a depth similar to those in the MM, and the DM since their minimum values were next to 0, whereas the sunspot number minimum value observed in the GM was slightly above (around 15) considering data until 1894. However, the duration of the SM was shorter than the MM and DM (as in the case of the aurora series), and similar to that occurred in the GM. In this case, sunspot number values below the threshold used in Section 4.1 for the SN_{14C} series (~ 20), spanned the period of 1748–1752 for the SM, whereas they were observed during all the years of the MM (1645–1715), 1796–1820 for the DM, and 1883–1887 for the GM.

The SM is not present as a remarkable minimum in the SN series. There is a decrease in the sunspot number around the 1750s to values as low as ~ 65 . However, the depth of this relative minimum is significantly different from those of the MM and DM since sunspot number values in those minima were around 25 in the last part of the MM (which is the only available) and the DM. The depth of the SM is merely comparable to that of the GM when sunspot number values reached minimum values around 50. Moreover, comparing the time extension of both minima according to the SN series, values of the SN below the threshold used previously (~ 70) ranged from 1745 to 1759 for the SM and they cover the period 1795–1830 for the DM and 1876–1918 for the GM.

4.3. Can a non-remarkable minimum in the SN series be compatible with a remarkable minimum in another solar activity indices?

4.3.1. Evidence from the Silverman Minimum

The SM appears as a pronounced minimum both in the AD and SN_{14C} series in terms of depth and also in duration in the case of the AD series. However, considering the SN series, the SM is a weak non-remarkable minimum only comparable in depth to that of the GM rather than to periods of very reduced solar activity such as the DM and the MM (which is the only grand minimum observed in the telescopic era).

Interestingly, a similar behavior is observed in the GM, which shows a notable minimum in AD, but only a relatively weak minimum in SN. Furthermore, the SM and GM present comparable minimum values in both AD and SN. Therefore, this suggests that a remarkable minimum in AD could be compatible with a non-remarkable minimum in SN.

The minimum periods for which the SM occurs in each series are similar, although they do not overlap exactly. However, partial temporal offsets are also observed in the other studied minima (MM, DM and GM), being a clear example the end of the MM. Therefore, the differences found for the SM are not exceptional within the context of long-term solar variability.

Another issue particularly interesting is that, unlike the GM, the SM does not occur near the beginning or end of a centennial Gleissberg cycle (Gleissberg, 1939, 1945), where reduced solar activity is more commonly observed, but around its middle phase.

4.3.2. Could the discrepancies arise from data limitations?

Part of the discrepancies among the three proxies may arise from their associated uncertainties and observational limitations.

The uncertainty associated with the SN values provided by SILSO is only available from 1818 onwards. However, an important limitation of the SN series is the sparse temporal coverage of historical sunspot observations around the SM.

First, the annual number of sunspot observations used to compute the SN series is not available from SILSO for the 18th century. However, using as reference an updated version of the database of Vaquero *et al.* (2016) included in Clette *et al.* (2023), one can check that the temporal coverage of sunspot observations for the decades 1720s–1810s is low (see Table 2): between 3 % in the 1730s and 37 % in the 1770s. Second, the temporal coverage from SILSO data can be calculated from the 1820s onwards (Table 2), revealing substantially more complete observations during later periods, including a temporal coverage of 100 % from the 1850s.

Therefore, a significant fraction of the daily sunspot observations around the SM is missing, meaning that the SN series for that period is less reliable and may be more difficult to characterize than for later periods. Some works have been carried out to analyze and to incorporate new sunspot records in periods with scarce temporal coverage (Carrasco, Villalba-Álvarez, and Vaquero, 2015; Nogales *et al.*, 2020; Hayakawa *et al.*, 2022). However, much work is still needed to improve the reconstruction of solar activity during the mid-18th century.

In relation to AD, uncertainties are more difficult to quantify because the index is derived from the catalog including a list of dates when auroras were observed. Nevertheless, the catalog compiled by Křivský and Pejml (1988) was made up of a collection of numerous catalogs and observations, including quality criteria to ensure that the data series was as robust as possible. For example, an aurora record was incorporated into the catalog only if the aurora was observed simultaneously at several locations separated by a considerable distance apart.

One could also think that the SM is the result of a mischaracterization of AD during that period. For example, the Seven Years' war took place during 1756-1763 and could have affected the record of auroras. However, the SM was beyond the end of the war. In fact, the AD series is more affected by solar activity than by the numerous wars around the world as demonstrated by the AD minima coinciding with the SN minima (in both 11-year cycles and Gleissberg cycles, see Figure 1 and 2).

In the case of SN_{14C} , the average 1-sigma uncertainty during the SM is approximately 15 units in sunspot number. Although this uncertainty is not negligible, the minimum observed in the

SN_{14C} series is sufficiently pronounced to remain identifiable despite that uncertainty. Nevertheless, the uncertainty may partially affect the exact depth and duration of the minimum.

4.3.3. Physical interpretations and implications

The solar phenomena responsible for geomagnetic disturbances, which result in significant aurorae, are CME and HSS. Around solar cycle maxima, when SN reaches its highest values, the CME occurrence rate peaks (Webb, 1991), making CMEs the dominant driver of geomagnetic storms (Richardson, Cliver, and Cane, 2001). In contrast, HSSs become more important near solar cycle minima (Gonzalez, Tsurutani, and Clúa De Gonzalez, 1999). While CMEs can cause very intense geomagnetic activity and bright auroras extending toward mid and low latitudes, auroras associated with HSSs, although more numerous, are generally weaker and only the fastest HSSs can produce mid-latitude auroras.

Figures 1 and 2 (upper panels) show that higher 11-year SN cycles are generally associated with higher AD values, whereas weaker SN cycles have greater variability in AD values. Thus, cycles of similar amplitude in SN may present significant differences in AD values, particularly during weak cycles, such as those around the SM. Apart from the uncertainties associated with the time series commented above, this behavior can also contribute to explain the difference in the minimum periods identified for the SM in the AD, SN and SN_{14C} series.

The solar phenomena causing mid-latitude auroras could be behind the above behavior. During the maximum of stronger cycles,

Table 2. Temporal coverage of sunspot observations per decade for the period 1700-1819 according to an updated version of the database of Vaquero *et al.* (2016) used by Clette *et al.* (2023) and the period 1820-1949 according to SILSO.

DECADE	COVERAGE (%)	DECADE	COVERAGE (%)
1700	73.3	1830	62.0
1710	68.2	1840	76.5
1720	27.3	1850	100
1730	3.1	1860	100
1740	5.9	1870	100
1750	35.5	1880	100
1760	32.0	1890	100
1770	37.2	1900	100
1780	5.6	1910	100
1790	23.8	1920	100
1800	26.0	1930	100
1810	60.2	1940	100
1820	79.9		

the Sun tends to develop numerous magnetically complex active regions produce frequent and fast CMEs, that cause an increase in the number of mid-latitude auroras. In weaker cycles, fewer and less magnetically complex active regions are created. Since the SN index depends on the total number of spots but also on the number of sunspot groups (Clette *et al.*, 2023), similar SN values may correspond to substantially different magnetic configurations. In particular, a given SN value can be obtained with dispersed and sufficiently complex active regions that can produce CMEs, or with less disperse and magnetically complex active regions that can produce CMEs (Lamy *et al.*, 2019). Thus, low-amplitude SN cycles tend to yield fewer and more variable mid-latitude auroras.

Weaker solar magnetic fields are associated with a higher number of near-equatorial coronal holes, which produce strong and recurring HSSs (Gibson *et al.*, 2009). Typically, these streams generate geomagnetic storms leading to auroras visible only at high terrestrial latitudes, but some of the strongest streams may cause mid-latitude auroras. This can be seen in Figures 1 and 2 (upper panels) since relatively high AD values can be found in some periods of low SN near solar minima. Thus, for low amplitude cycles in SN, HSSs may be more likely to contribute to generating mid-latitude auroras resulting in a more variable number of mid-latitude auroras. That is, cycles of low amplitude in SN could tend to produce a more variable number of mid-latitude auroras through both CMEs and HSSs.

These phenomena can occur more frequently at close intervals. Roughly CMEs peak at solar maxima and HSSs at solar minima. Some studies even indicate that CME occurrence rate tends to have two peaks: one around solar maximum and the other 1-3 years after solar maximum (Gonzalez, Tsurutani, and Clúa De Gonzalez, 1999; Robbrecht, Berghmans, and Van der Linden, 2009). Meanwhile, HSSs activity peaks during the late declining phase of the solar cycle, when coronal holes extend from the solar polar regions down to the equator (Mursula and Zeiger, 1996; Borovsky and Denton, 2006). Therefore, the geomagnetic activity often increases during the declining phase of the solar cycle (Legrand and Simon, 1991). This is consistent with Figures 1 and 2 (upper panels) where many AD maxima peak in the declining phases of SN cycles. This highlights the fact that the mechanism of aurora production makes the relationship between AD and SN intrinsically variable, especially in the short term, whereas the relationship between the three studied series becomes more similar in the long term (Figure 1 and 2, lower panels).

5. Conclusions

Silverman (1992) stated that a prolonged minimum around 1765, previously unrecognized, is evident from the series of

annual number of days on which northern auroras were sighted in mid-latitudes (below 55°N in geographic latitude). In the present work, we have compared this type of series (AD) with the solar activity level according to the sunspot number (version 2) provided by SILSO (SN) and the more recently published high-resolution series of annual sunspot number derived from ¹⁴C concentrations in tree rings (SN_{14C}) by Usoskin *et al.* (2021) for the period 1600-1950, highlighting the 18th century with the goal of assessing whether the Silverman Minimum can be considered a remarkable minimum in solar activity or is only a minimum in aurora observations.

The three data series show similarities in their time evolutions, especially in the long term. A pronounced minimum is found in AD (period 1755–1766) and SN_{14C} (1747–1752) series with a depth similar to those of the MM and DM, whereas SN series just shows a weak non-remarkable minimum (1745–1761) with a depth merely comparable to that of the Gleissberg minimum occurred at the end of 19th century and early 20th century.

A mischaracterization of AD could be argued to explain the discrepancy. However, the pronounced minimum is not only found in AD, but also in SN_{14C}. In addition, the time evolution of AD is in line with 11-year and centennial cycles and with a geomagnetic behavior that can be explained in terms of solar phenomena such as CMEs and HSSs rather than non-physical factors, such as wars, thanks to the use of a catalog of auroras made up of numerous catalogs with corrections and criteria so that the data series was as robust as possible. A mischaracterization of SN could be argued due to the lower temporal coverage of this index in the 18th century. However, at the end of the 19th century, there is a similar non-pronounced minimum in SN (which is characterized with 100 % temporal coverage) with a notable minimum in AD and SN_{14C} series. This proves that a prolonged low solar activity (without being a pronounced minimum) may produce a remarkable minimum in other solar activity indices (as is the case of the Silverman Minimum in aurora records in the mid-18th century) without implying that any of the series are mischaracterized.

We highlight that the SN series is the best index to characterize long-term solar activity. However, it does not provide enough information to describe the magnetic complexity of the heliosphere and its impact on our planet (Gibson *et al.*, 2009). Nevertheless, more studies incorporating new sunspot records made in the mid-18th century to the databases are necessary to definitely clarify the true difference in the three series considered in this work so that similarities and differences between solar and geomagnetic activity become clearer. Thus, taking into account these similarities and differences, SN can be useful as a first approximation in estimations of geomagnetic behavior on large temporal scales for periods with no or scarce geomagnetic data.

Finally, we would like to pay tribute to Prof. Sam Silverman within the context of this work. He peacefully passed away at his home on April 23, 2022, at the remarkable age of 96. A physicist and lawyer by profession, his enduring curiosity led him to publish across a multitude of disciplines, ranging from his primary focus on upper atmospheric physics and the aurora borealis to a diverse array of subjects in the social sciences. Within the meteorology and space weather community, he will forever be commemorated as one of the foremost researchers in airglow and auroras globally. Arguably, his most renowned paper delved into the secular variation of the aurora over the past five centuries (Silverman, 1992), although his prolific body of work widely extended to historical observations of this phenomenon. His collaborations with distinguished members in the space weather and climatology realm, such as Prof. Joan Feynman (Feynman and Silverman, 1980; Feynman *et al.*, 1990a, 1990b), Prof. Ralph Shapiro (Silverman and Shapiro, 1983), and Prof. Ed Cliver (Silverman and Cliver, 2001), further solidified his impact. His most recent publication, achieved in partnership with Prof. Hisashi Hayakawa, contributed to the study of Dalton's auroral observations (Silverman and Hayakawa, 2021), marking yet another testament to his enduring dedication and influential legacy.

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7. Author contributions

JMV had the idea of the manuscript. VMS and AJP Aparicio made the calculations included in the manuscript and wrote the first version. All the authors discussed the results and reviewed the manuscript.

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9. Data Availability

The data used in this work are available in the original references mentioned in the text. The authors can provide these data upon request.

10. Competing interests

The authors declare no competing interests.

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The background of the page is a light orange color. It features a network of thin, curved orange lines that connect various circles of different sizes. The circles are also orange and vary in diameter, with some being significantly larger than others. The overall pattern is organic and interconnected, resembling a molecular structure or a network diagram. The text is placed within white rectangular boxes that have a slight drop shadow, making them stand out from the background.

Sección especial

Estudios Métricos en
Ciencias de la Tierra

Special section

Metric Studies in
Earth Sciences

Las erupciones del volcán Irazú (Costa Rica) entre 1924 y 1940 a partir de una reconstrucción desde los discursos hemerográficos

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Abstract

Over the past half-century, numerous eruptions of the Irazú volcano have frequently been cited as occurring between its two main historical eruptive cycles in 1723–1724 and 1963–1965, particularly during the 1920s and 1930s. To date, no eruptions have been documented at Irazú during the 19th century, and a reliable chronology was recently reconstructed for the eruptive period between 1917 and 1921, the only one documented after that of 1723–1724. This study involved a review of newspaper archives and scientific sources regarding the alleged eruptions between 1921 and 1940, which established that during those decades, eruptions occurred only in June 1933 and between September 1939 and April 1940. Several of the alleged eruptions have been misinterpretations based on isolated and incomplete sources in association with strong seismic events, such as that of March 1924, or with intense exhalative activity that extended from 1921, which produced high columns of steam from the summit craters of Irazú and were generally called “eruptions,” without actually being so. The ash produced in 1933 reached the capital, San José, and is estimated to have had a combined VEI of 2. On the other side, 1939–1940 eruptions were smaller in size, although more spread out over time, and the ash fall towards the Western Central Valley was much lighter. The damage and overall impact of both eruptions were of little economic and social significance, and although the government acted in isolation to address the crises, there was no need for major interventions. This new clarification of the correct historical eruptive record of Irazú will allow for a better assessment of the hazard and eruptive frequency, in the interest of better and more efficient risk management in the affected areas.

Key words: Irazú volcano, historic eruption, volcanic chronicle, eruptive history, volcanic hazard.

Resumen

A lo largo del último medio siglo ha sido frecuente citar múltiples erupciones del volcán Irazú entre sus dos ciclos eruptivos históricos principales en 1723–1724 y 1963–65, en particular en las décadas de 1920 y 1930. Hasta el momento no se han documentado erupciones durante el siglo XIX en el Irazú y recientemente se reconstruyó una buena cronología para el periodo eruptivo entre 1917 y 1921, el único posterior documentado al de 1723–1724. En este trabajo se realizó un estudio hemerográfico y de fuentes científicas sobre las supuestas erupciones entre 1921–1940 que permitió establecer que en esas décadas solo hubo erupciones en junio de 1933 y entre setiembre de 1939 y abril de 1940. Varias de las supuestas erupciones han sido malinterpretaciones a partir de fuentes aisladas e incompletas en asociación con fuertes eventos sísmicos, como el de marzo de 1924, o bien con una intensa actividad exhalativa que se extendió desde 1921, que generaba columnas de vapor desde los cráteres cuspidales del Irazú y en general fueron llamadas “erupciones”, sin serlo. Las cenizas producidas en 1933 alcanzaron la capital San José, y se estima que en conjunto tuvo un equivalente a un VEI de 2. En cambio, las erupciones de 1939–1940 fueron más discretas en tamaño, aunque más esparcidas en el tiempo eruptivo, y la caída de ceniza hacia el Valle Central Occidental fue mucho más tenue. Los daños y afectación en general de ambas erupciones fueron de poca monta económica y social, y aunque el gobierno actuó de forma aislada para atender las crisis, no hubo necesidad de grandes intervenciones. Esta nueva clarificación del correcto registro eruptivo histórico del Irazú, permitirá una mejor evaluación de la amenaza y frecuencia eruptiva, en aras de una mejor y más eficiente gestión del riesgo en las áreas afectables.

Palabras clave: volcán Irazú, erupción histórica, crónica volcánica, historia eruptiva, peligro volcánico.

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1. Introducción

El volcán Irazú (9,979°N y 83,852° O, 3432 m s.n.m) es un amplio complejo estratovolcánico andesítico basáltico (excepcionalmente también presenta dacitas) situado en el extremo sudoriental de la cordillera volcánica Central (CVC) de Costa Rica (Alvarado, 1993; Alvarado *et al.*, 2006; Clark *et al.*, 2006). Esta cordillera forma, junto con las cordilleras volcánicas de Guanacaste (CVG) y de Tilarán (CVT), el arco volcánico cuaternario de Costa Rica, el cual se origina por la subducción de la placa del Coco bajo la placa del Caribe y cuyo eje se encuentra unos 150 km tierra adentro de la Fosa Mesoamericana, con una orientación noroeste-sureste (Figura 1a). El volcán Irazú se localiza unos 25 km al este del centro de la ciudad de San José, capital de Costa Rica y 15 km al noreste de la ciudad de Cartago. Ambas conforman, actualmente, junto con las ciudades de Heredia y Alajuela, la Gran Área Metropolitana (GAM), el centro socio-económico y más densamente poblado de Costa Rica, lo que configura un escenario de riesgo volcánico relevante (Soto y Sjöbohm, 2015; Campos-Durán *et al.*, 2024b) (Figura 1c).

Las principales erupciones históricas del Irazú ocurrieron en 1723-24 y quizás hasta 1726 (estromboliana, freatomagmática y freática, Alvarado y Schmincke, 2013), 1917-1921 (freática, freatomagmática-vulcaniana, estromboliana; Mora *et al.*, 2024) y 1963-65 (freática, freatomagmática y estromboliana a vulcaniana; Murata *et al.*, 1966; Barquero 1975; Alvarado 1993). Otras fases menores se han reportado entre 1924 y 1940 (Dóndoli, 1965; Barquero, 1975, 1976; Alvarado 1993), aunque no se han investigado a fondo, por lo que no se conocen muy bien y son el objeto de este estudio. Luego de la última fase eruptiva de 1963-1965, el Irazú ha mantenido una tenue actividad fumarólica en el flanco norte, muy cerca de la cima, con aparición y desaparición periódica con variaciones del nivel, color, temperatura y acidez de una laguna intracrática, enjambres sísmicos ocasionales y la generación de grandes deslizamientos en el flanco norte superior del macizo. Uno de estos eventos se asoció con una erupción freática en 1994 al noreste del cráter principal, en el sitio conocido en el siglo XIX y principios del XX como “volcán Nuevo” (Fallas *et al.*, 2018) (Figura 1b). El conocimiento histórico y prehistórico de la actividad eruptiva explosiva del Irazú (p. ej., Clark *et al.*, 2006; Campos-Durán *et al.*, 2024a, 2024b) permite sugerir que las erupciones magmáticas-freatomagmáticas con $VEI \geq 3$ suceden con periodos de retorno del orden de 200-400 años, mientras que los episodios con $VEI < 3$ suceden con recurrencias pluridecadales (Soto y Sjöbohm, 2015), pero luego de periodos de calma eruptiva de varias décadas posteriores a las paroxísmicas, como fue el caso pos 1723-26 y es el caso pos 1963-65.

En este trabajo se analiza la información disponible sobre reportes de actividad del volcán Irazú entre 1924 y 1940, con

el fin de establecer las fases eruptivas que habrían ocurrido en ese intervalo de tiempo y su impacto, de modo que se constituya en un insumo fundamental para los análisis de riesgo volcánico y como antecedente para la prevención de desastres de origen volcánico en Costa Rica. El enfoque desde la historia y la vulcanología permiten la generación de conocimiento para completar el registro histórico de actividad de un volcán, en el contexto de los hechos políticos, económicos y sociales de la época en la que ocurrió la actividad eruptiva de interés, tal y como fue abordado por Mora *et al.* (2024) para la fase eruptiva de 1917 a 1921 del volcán Irazú.

2. Método

2.1. Fuentes de información y curación de datos

El análisis se realizó con base en fuentes hemerográficas que se encuentran disponibles, en forma digital y física (según su estado de preservación), en la Hemeroteca de la Biblioteca Nacional de Costa Rica, accesible digitalmente desde el portal del Sistema Nacional de Bibliotecas (SINABI, <https://www.sinabi.go.cr/>). Se analizaron 187 noticias que fueron transcritas y sistematizadas cronológicamente en una base de datos organizada con los siguientes elementos: periódico, página, autoría, título de la noticia, día, mes, año y la transcripción. Las transcripciones de las noticias mantienen, en gran medida, los rasgos gramaticales y ortográficos originales; solo se corrigieron algunos errores tipográficos de imprenta.

Los textos relativos a la actividad eruptiva y su impacto se extrajeron manualmente y se analizaron desde el prisma de la percepción, la cual es permeada por el contexto cultural e histórico del momento, así como también en contraposición y complemento por la convergencia de otros procesos naturales como los sismos, que para la primera mitad del siglo XX aún eran cercanamente correlacionados según la interpretación no científica, y particularmente la noticiosa. La comparación de los discursos contenidos en las noticias analizadas permite identificar sesgos y distorsiones y resalta los elementos consistentes sobre la actividad volcánica. Finalmente, esta información se calibra, al menos en parte, con la información científica de la misma época obtenida no solo de las fuentes hemerográficas, sino también del Fondo José Fidel Tristán (FTRIS) que forma parte del Archivo Histórico (AH) del Archivo Nacional de Costa Rica (ANCR), accesible desde su portal (<https://archivodigital.go.cr/>). Finalmente, se interpretan los hechos a la luz del conocimiento científico actual, lo que obliga a generar hipótesis y suposiciones con respecto a la información localizada (Mora *et al.*, 2024). La contextualización histórica se basó en periódicos, revistas, obras

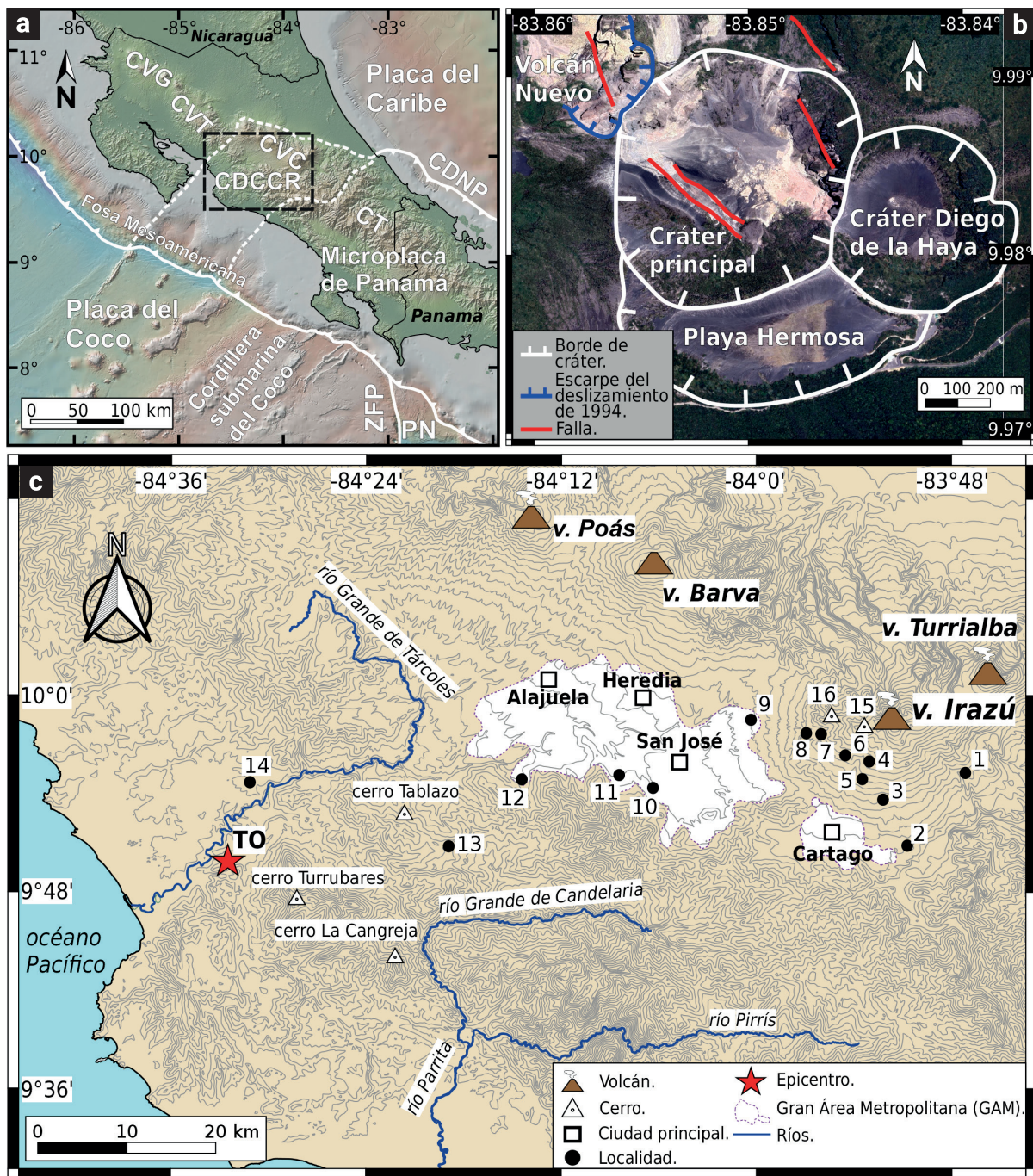


Figura 1. a) Contexto tectónico simplificado de Costa Rica. Abreviaturas: cordillera Volcánica de Guanacaste (CVG), cordillera Volcánica de Tilarán (CVT), cordillera Volcánica Central (CVC), cordillera de Talamanca (CT), Cinturón Deformado del Centro de Costa Rica (CDCCR, región dentro de la línea punteada con base en Denyer *et al.*, 2013), Cinturón Deformado del Norte de Panamá (CDNP), Zona de Fractura de Panamá (ZFP) y la placa Nazca (PN). Los triángulos indican movimiento inverso y apuntan hacia la placa cabalgante. La batimetría y topografía se obtuvieron de Ryan *et al.* (2009). El rectángulo en línea discontinua de color gris oscuro representa el área ampliada que se muestra en c. b) Detalle de la cima del volcán Irazú con base en una imagen de Google Earth CNES/Airbus, Maxar Technologies de abril de 2023 (Modificada de Mora *et al.*, 2024). Las fallas se indican según el trabajo de Ulloa *et al.* (2018). c) Detalle del área ampliada señalada en a). Se muestra la ubicación de las ciudades principales, localidades, ríos y elementos geográficos mencionados en el texto. El mapa y curvas de nivel están basados en el modelo de elevación digital suministrado por el Instituto Geográfico Nacional (IGN). Los números indican las siguientes localidades y rasgos: 1) Capellades, 2) Birrisito, 3) Cot, 4) Sanatorio Carlos Durán, 5) Tierra Blanca, 6) Llano Grande, 7) Corralillo (actual Hacienda Guayabillos), 8) Rancho Redondo, 9) San Isidro de Coronado, 10) Alajuelita, 11) Escazú, 12) Villa Colón (actual Ciudad Colón), 13) Puriscal, 14) Orotina, 15) Cerro Cabeza de Vaca, 16) Cerro Pico de Piedra (ver figuras 2 y 4). TO: Epicentro del terremoto de Orotina con base en Montero (1999).

literarias y publicaciones relativas al contexto histórico, social y económico de la época, accesibles desde el portal del Sistema Nacional de Bibliotecas de Costa Rica (SINABI), bibliotecas institucionales y colecciones privadas.

2.2. Definición de los periodos de análisis

En 1924, la supuesta actividad eruptiva del volcán Irazú fue vinculada por la prensa escrita al sismo del 4 de marzo de ese año, conocido como el terremoto de Orotina o de San Casimiro, con epicentro aproximadamente 22 km al suroeste de Orotina, una profundidad menor a 40 km y una magnitud $M_s = 7,0$ (Montero, 1999; Figura 1c). Las réplicas se extendieron durante todo marzo y parte de abril, acompañadas por una intensa sismicidad inducida en otras fuentes corticales (Montero, 1999). El evento causó daños severos tanto en el Pacífico Central como en el Valle Central, particularmente en San José, lo que generó una amplia cobertura mediática en la prensa escrita, con más de 300 noticias publicadas. De estas, únicamente 22 hacen referencia a la supuesta actividad eruptiva del Irazú y son las que se utilizan en este trabajo.

Entre 1925 y 1932 se identificaron numerosos reportes de supuesta actividad eruptiva en el volcán Irazú, muchos en un único periódico (Figura 2). En contraste, en 1933 y de 1939 a 1940 se dispone de un conjunto más consistente de noticias que documentan caídas de ceniza y algunos impactos asociados. Para el periodo 1934-1938 no se registraron referencias a actividad eruptiva; únicamente se identificaron dos noticias que aluden a supuestas erupciones en el volcán Poás (Figura 2).

Con el fin de cuantificar lo anterior, se analizaron las 187 noticias mediante la búsqueda y contabilización de tres palabras clave: *irazú*, *erupción* y *ceniza*. Las dos primeras aparecen en la mayoría de los registros, mientras que la mención de *ceniza* se concentra principalmente en los años 1933, 1939 y 1940; su presencia es escasa en 1924 y 1932, y está ausente entre 1935 y 1938 (Figura 3). De modo que, con base en la frecuencia de reportes, consistencia entre fuentes y presencia de indicadores directos de actividad eruptiva (p. ej. caída de ceniza), el análisis y la presentación de los resultados se organizó en cuatro periodos: 1924, 1925-1932, 1933 y 1939-1940 (Figura 2).

3. Breve contexto histórico-científico de la época

A finales del siglo XIX y durante la primera mitad del siglo XX la vulcanología en Costa Rica no fue favorecida institucionalmente y su desarrollo recayó en un grupo de científicos de la época que se encargaron, con poco tiempo y recursos, de explorar y describir los volcanes y su actividad, así como

otros relieves, rocas y procesos ígneos (Mora, 2024). Una de las voces predominantes en el ámbito del estudio de volcanes fue la de José Fidel Tristán, quién describió varias erupciones volcánicas, entre ellas, las erupciones de 1917 a 1921 del volcán Irazú (Mora *et al.*, 2024). También, participó activamente en las discusiones sobre la actividad sísmica ocurrida en 1924 y su impacto (Mora, 2024). Con la muerte de José Fidel Tristán, en 1932, hubo un cambio hacia otros científicos geólogos que llegaron del exterior, como Paul Schaufelberger (1929 a 1938) y de César Dóndoli (1939 en adelante) (Mora, 2024). Otras voces fueron las del Ing. Rafael Tristán, hermano de José Fidel, así como también las de Ricardo Fernández Peralta, Anastasio Alfaro, Rubén Torres Rojas, Carlos Borel, Manuel Valerio, Fernando Torres Vincenzi, entre otros; y aunque ninguno tuvo el rol protagónico que tuvo José Fidel Tristán en el ámbito vulcanológico, dejaron valiosos relatos y descripciones sobre el Irazú en la prensa escrita que permiten sustentar el análisis sobre la actividad del Irazú en este estudio.

4. Vínculo de las erupciones volcánicas y actividad sísmica desde la percepción

El vulcanismo y la sismicidad, desde el prisma de la percepción, son dos procesos geológicos que producen un gran asombro y temor en el ser humano, ya sea por el evento mismo (evidentemente disruptivo de la cotidianidad), la tradición, la religión, las creencias, las leyendas, la literatura o el cine, de modo que, quienes conviven con los procesos volcánicos, han llegado a tejer en su imaginario individual y colectivo, gran cantidad de explicaciones no formales y saberes populares en torno a la actividad volcánica y sísmica (Peraldo y Mora, 2009, 2016). Peraldo y Mora (2009) estudiaron el miedo inducido por la prensa sensacionalista, la cual genera noticias en torno a eventos naturales carentes de sustento científico ni corroboración, y que por tanto exacerbaban los ánimos ya caldeados por la actividad sísmica. Esto resulta en la percepción de una relación directa, causa-efecto, entre los sismos y las erupciones volcánicas, tal y como ocurrió con varios sismos importantes que ocurrieron entre 1924 y 1940 y que causaron gran consternación sobre si eran causantes de erupciones volcánicas o viceversa (Tabla 1 del [Apéndice A](#)).

Por otro lado, algunas interpretaciones de la comunidad científica de la época se alineaban con la percepción general de la relación directa sismos-erupciones. Uno de ellos fue el Ing. Rafael Tristán (hermano de José Fidel Tristán, fallecido en enero de 1932) quien asoció, por ejemplo, los temblores ocurridos el 13 de julio de 1933 (La Tribuna, 1933c) y un sismo ocurrido el 18 de junio de 1939 (La Prensa Libre, 1939) con la actividad

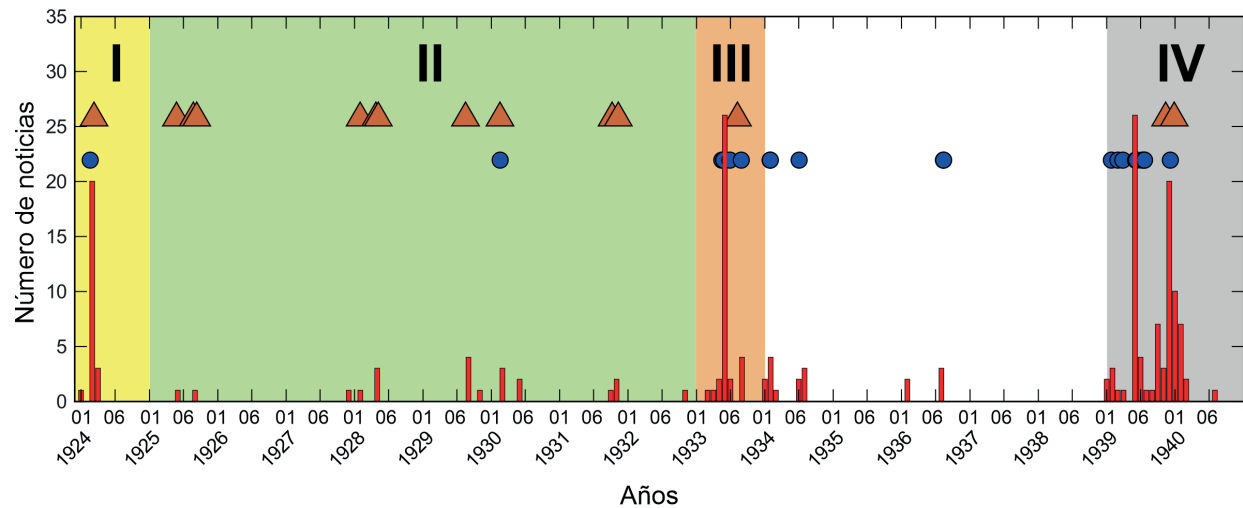


Figura 2. Número de noticias que reportan actividad eruptiva del volcán Irazú entre 1924 y 1940. Los triángulos indican supuesta actividad eruptiva reportada. Los círculos azules indican los sismos reportados en la prensa. Los colores de fondo y los números romanos señalan los periodos en los que se organizó el análisis en este estudio.

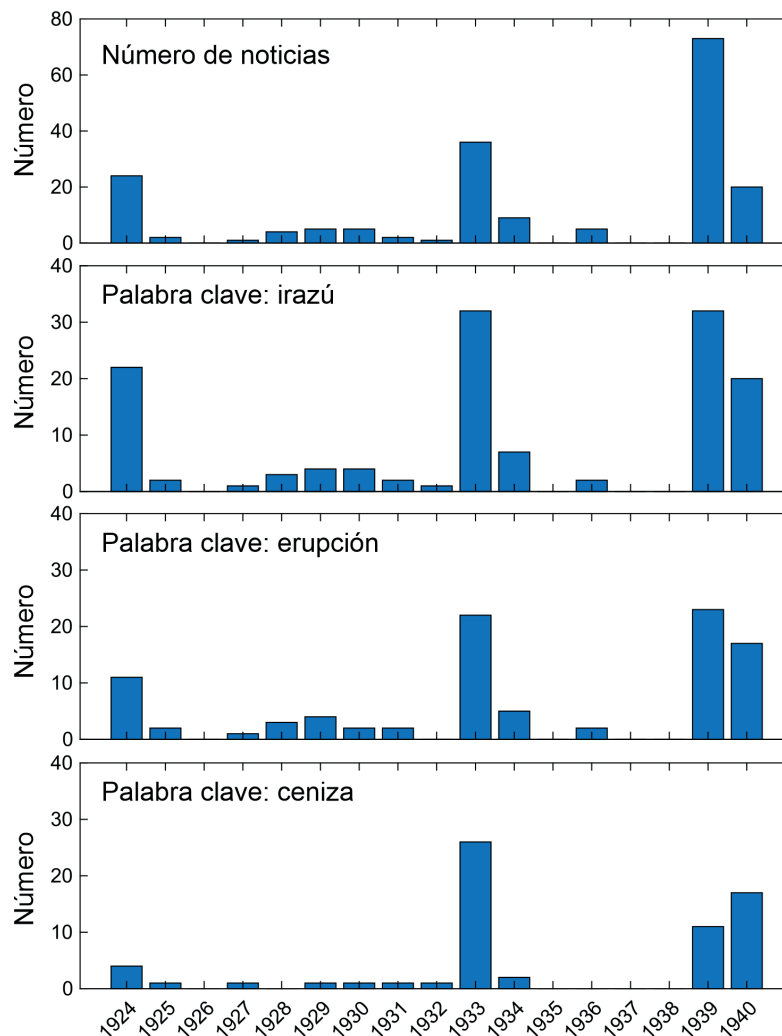


Figura 3. Frecuencia anual de noticias (arriba) y de las palabras clave *irazú*, *erupción* y *ceniza* (paneles inferiores) en el conjunto de 187 noticias analizadas.

del volcán Irazú. Más interesante todavía es que el mismo Ing. Rafael Tristán primero manifestó que la sismicidad producto del terremoto de Puerto Armuelles del 17 de julio de 1934, que se extendió por varios días, estaba asociada a una falla (La Prensa Libre, 1934a) y luego, en otra entrevista, contradictoriamente sugirió un origen volcánico (La Prensa Libre, 1934b). Esta dualidad pone en evidencia el pobre entendimiento del origen de los procesos sísmicos y volcánicos, así como también del concepto de falla tectónica activa.

Unos años más tarde, Paul Schaufelberger, geólogo que estuvo en el escenario científico de Costa Rica entre 1929 y 1938 (Mora, 2024), aunque no relacionó los temblores ocurridos el 13 de julio de 1933 con la actividad del volcán Irazú, sí interpretó que se habrían generado por “hundimientos de cavernas subterráneas” producto de la acción química y mecánica del agua subterránea (La Tribuna, 1933d). En contraste, el mismo Paul Schaufelberger atribuyó un origen volcánico a los sismos ocurridos el 26 y 27 de agosto de 1936, tal y como lo hizo la prensa (La Prensa Libre, 1936a, 1936b).

En esa época, una idea análoga a las “cavernas de aguas subterráneas” vinculaba los sismos con la actividad volcánica a través de la formación de cavidades bajo los volcanes como resultado de la gran cantidad de ceniza que era lanzada y que debían “acomodarse” en algún momento, tal y como se pensó que ocurrió durante la erupción del volcán Poás que dejó una cavidad que luego originó los terremotos en 1910 (p. ej. Serre del Sagués, 1920). De forma similar, se pensó que la fase eruptiva de 1917 a 1921 del Irazú debió formar grandes cavidades que originaban los movimientos sísmicos que se sintieron en 1924, lo cual fue retomado por Paul Schaufelberger para interpretar el origen de la actividad eruptiva del volcán Irazú en 1933 (Diario del Comercio, 1924c; La Hora, 1933; Schaufelberger y Jiménez, 1933). También hubo científicos que refutaban la idea de que los temblores se generaran por actividad volcánica, y uno de ellos fue José Merino y Coronado (Novedades, 1939b). No fue sino hasta 1939 que César Dóndoli introdujo, categóricamente, la idea del origen tectónico de los eventos sísmicos (Dóndoli, 1939).

5. Resultados

5.1. Primer periodo: 1924

5.1.1. El terremoto de Orotina y el volcán Irazú

Como consecuencia del terremoto de Orotina surgieron en los periódicos numerosas noticias imprecisas, e incluso contradictorias, sobre supuestas erupciones en el Irazú (Tabla 2 del [Apéndice A](#)). El propio día del temblor (4 de marzo), mientras

La Noticia y *La Tribuna* informaron que el Irazú se encontraba tranquilo, *La Prensa* y *el Diario de Costa Rica* reportaron que sí había hecho erupción (Tabla 2 del [Apéndice A](#)). El 5 de marzo, *La Voz*, *La Prensa* y *La Noticia* informaron que desde Tierra Blanca (Figura 1c) se escuchaban retumbos, se veían relámpagos, llamas y erupciones de humo denso y negro; mientras que el *Diario de Costa Rica* refería que esas observaciones eran falsas y exageradas e informó que tanto Tierra Blanca como Cartago estaban tranquilos (Tabla 2 del [Apéndice A](#)). La *Nueva Prensa* reportó, el 5 de marzo, un “penacho blanco” que catalogó como erupción, y el día siguiente (6 de marzo), informó que el Irazú no presentaba ninguna actividad especial como tampoco ceniza ni lodo en las erupciones (Tabla 2 del [Apéndice A](#)).

Entre el 7 y el 9 de marzo de 1924 surgió una polémica entre Francisco Jiménez, entonces Gobernador de Cartago, y el periódico *La Tribuna* (Tabla 2 del [Apéndice A](#)). El primero, con base en informes del Lic. Arturo Volio, desmentía que el Irazú estuviera arrojando cenizas, en tanto que el segundo respondió con un único reporte de una erupción de ceniza en poca cantidad desde un caserío de Llano Grande (Figura 1c). El 9 de marzo el periódico *La Verdad* hizo eco de ese reporte de caída de ceniza, mientras que, el 11 de marzo *La Tribuna* publicó, según su corresponsal en Cartago, que el Irazú “no daba señales de amenazadora actividad” y solo se divisaba una “débil columna de humo” (Tabla 2 del [Apéndice A](#)). El 22 de marzo el *Diario del Comercio* informó sobre una erupción cuya ceniza fue dispersada hacia el sureste, pero es una noticia aislada y no es posible confirmarla (Tabla 2 del [Apéndice A](#)). A finales de marzo y también en abril hay algunas referencias de que el Irazú presentaba actividad, pero no son precisas en describir erupciones propiamente dichas (Tabla 2 del [Apéndice A](#)). Lo que sí es consistente, y llama la atención, es la ausencia de reportes explícitos de caída de ceniza cerca o allende del cráter activo o de algún impacto producto de la actividad eruptiva (Tabla 2 del [Apéndice A](#)).

El científico y profesor José Fidel Tristán (en ese entonces director del Liceo de Costa Rica) realizó observaciones diarias de la actividad del volcán Irazú entre 1924 y 1927 desde la ciudad de San José y, ocasionalmente, desde Rancho Redondo de Goicoechea (Figura 1c), al oeste del volcán, en un lugar llamado en ese entonces Corralillo, actual Hacienda Guayabillos (Tristán 1927a, 1927b; Figura 1c). La mayoría fueron hechas entre las 6:00 y las 8:00 de la mañana, pero también, ocasionalmente, las hacía en otros momentos del día. Aunque los reportes son pocos, evidencian que entre enero y febrero de 1924 el volcán Irazú presentaba una actividad exhalativa débil a moderada y un poco más conspicua en marzo de ese año. También reporta algunas erupciones, pero no hay descripciones explícitas que confirmen si hubo emisión de ceniza y/o caída de ella en alguna

parte, o bien, si solo fueron plumas de vapor de agua y gases (Tabla 3 del [Apéndice A](#)). A juzgar por las descripciones, una erupción para José Fidel Tristán se trataba de una salida de vapor o de gases, y no implicaba ceniza a menos que explícitamente haya escrito “erupción de ceniza”.

El 17 de marzo de 1924 llegó a Costa Rica el geólogo alemán Karl Sapper, quien ya había realizado una breve visita a Costa Rica en 1899, como parte de sus investigaciones geológicas de carácter regional en América Central (Sapper, 1942). De modo que el gobierno de Costa Rica aprovechó para incorporarlo a la comisión científica que ya había integrado, desde el 6 de marzo, con José Fidel Tristán, Ricardo Fernández y Anastasio Alfaro (Tristán, 1927a, folio 14; Mora, 2024). Como resultado, Karl Sapper publicó algunos trabajos científicos en los que hace algunas anotaciones relativas al volcán Irazú (Tabla 4 del [Apéndice A](#)).

Karl Sapper, con base en un reporte de Emilio Robert, hace referencia a una erupción grande, pero no dice la fecha en que ocurrió; aunque por contexto habría ocurrido el 4 de marzo a las 5:00 a. m., una hora después del terremoto de Orotina. Sin embargo, ya vimos que los periódicos publicaron información contradictoria al respecto y José Fidel Tristán no reporta ninguna actividad eruptiva relevante ese día (Tabla 3 del [Apéndice A](#)). El 5 de marzo, al amanecer, José Fidel Tristán describió varias erupciones a las que siguió un “desprendimiento constante de vapores”, congruente con lo descrito en el periódico *La Nueva Prensa* ese día (Tablas 2 y 3 del [Apéndice A](#)).

Karl Sapper menciona unas erupciones el 20 y el 23 de marzo en la madrugada, pero solo el *Diario del Comercio* reporta una “erupción de ceniza” en la noche del 20 y otras el 21 de marzo en la mañana (Tablas 2 y 4 del [Apéndice A](#)). Sin embargo, José Fidel Tristán reporta que el 21 de marzo en la mañana no había visto ninguna actividad (Tabla 3 del [Apéndice A](#)). Sobre la supuesta erupción del 23 de marzo resulta relevante que Karl Sapper, José Fidel Tristán y Ricardo Fernández Peralta se encontraban, ese día, hospedados en la casa de Emilio Robert (a pocos kilómetros de la cima del Irazú) y el 24 de marzo estaban en la cima del volcán. Por lo tanto, resulta extraño que no hicieran descripciones de la actividad eruptiva, por ejemplo, alguna caída de ceniza u otras evidencias que podrían observarse alrededor del cráter activo. Más aún, José Fidel Tristán solo reporta, lacónicamente, que el 23 de marzo a las 3:00 p. m. había ocurrido una erupción y en declaraciones que él brindó al *Diario de Costa Rica* del 26 de marzo, solo informa que las erupciones no son tan frecuentes como antes y que salen vapores blancos que se desvanecen rápidamente y dejan un “humo azul, que es ácido sulfúrico” (Tabla 2 y 3 del [Apéndice A](#)).

Sobre el ascenso al Irazú, Karl Sapper solo publicó una pequeña descripción sobre la actividad fumarólica en el cráter M (formado durante la fase eruptiva de 1917-1921, Fernández,

1920), muy similar a lo comentado por José Fidel Tristán en el *Diario de Costa Rica* el 26 de marzo de 1924 (Tablas 2 y 4 del [Apéndice A](#)). Por otra parte, en el libro publicado en 1925, *Los volcanes de la América Central*, que fue reproducido en la *Revista Conservadora del Pensamiento Centroamericano*, Sapper (1967) lista los eventos volcánicos ocurridos en América Central, y para 1924 no refiere ninguna actividad eruptiva del Irazú, solo menciona que el “4 de marzo comienza una serie larga de temblores en Costarica [sic]” (Sapper, 1967). Finalmente, en un manuscrito inédito titulado *Los efectos geográficos de los terremotos y erupciones volcánicas*, Karl Sapper hace una breve mención sobre unas nubes de polvo en el cerro Turrubares (Figura 1c) producto del terremoto de Orotina y no refiere nada sobre el Irazú (Sapper, 1924, folio 33).

Las descripciones lacónicas realizadas por José Fidel Tristán en 1924 sobre la actividad del volcán Irazú, contrastan con la amplia información que él y Ricardo Fernández Peralta brindaron durante la fase eruptiva de ese volcán entre 1917 y 1921 (Mora et al., 2024), lo que hace muy significativo que no detallaran las erupciones que supuestamente ocurrieron en marzo y abril de 1924. Karl Sapper en su visita a Costa Rica en 1924 se enfocó en otros temas, particularmente, en la erupción de 1723 del Irazú (Tristán, 1927b, folio 86) y lo poco que aportó sobre la actividad sísmica de 1924 se basó en lo que la comisión científica ya había concluido previo a su incorporación.

Adicionalmente, es muy sugestivo que algunas noticias sobre la ciudad de Cartago no mencionan nada sobre el volcán Irazú o su presunta actividad, solo mencionan lo poco que fue afectada por el terremoto de Orotina (*Diario de Costa Rica*, 1924c). El 16 de marzo de 1924 se comentó que el Irazú vivía “en su indiferencia de siempre”, refiriéndose a que no tenía nada que ver con el terremoto del 4 de marzo y el que no mencione nada explícito relativo al Irazú sugiere que el volcán no presentaba ninguna actividad significativa (*Diario de Costa Rica*, 1924e). De hecho, poco más de una década más tarde, en 1933, en el contexto de una fase eruptiva del Irazú que veremos más adelante, una noticia da a entender que no hubo ninguna actividad eruptiva entre la fase de 1917-1921 y ese año (*La República*, 1933). Finalmente, Dóndoli (1965, p. 5) y Murata et al. (1966) refieren que las únicas erupciones históricas bien documentadas, previas a la erupción de 1963-1965, fueron las fases mayores de 1723 y 1917-1921 y las fases menores de 1933 y 1939-1940.

5.1.2. La supuesta actividad volcánica del cerro Turrubares

Inmediatamente después del terremoto de Orotina, se informó que los cerros Turrubares (*Diario del Comercio*, 1924a y b; Figura 1c) y La Cangreja (*Diario de Costa Rica*, 1924b; Figura 1c) estaban en erupción. En los alrededores de Turrubares “el

vecindario está alarmado y sospecha que se trata del nacimiento de un nuevo volcán en Herradura o Turrubares” (Diario de Costa Rica, 1924a). La atención fue tal, que incluso se propuso que se trajeran dos aeroplanos desde Panamá para inspeccionar el cerro Turrubares y que sacaran fotografías que sirvieran a los científicos a entender lo que allí ocurría y que estaba alarmando a la población (Diario del Comercio, 1924a). Un testimonio describe que “la tierra se abrió [en la región de Turrubares] en diferentes lugares dejando enorme grietas, y que en algunos sitios se desprendían vapores sulfurosos y brotaba agua hirviendo” (Diario del Comercio, 1924b). También surgieron reportes sobre columnas de humo, un cráter e incluso de lava y fuego y se hizo el llamado a que se investigara para constatar si se trataba de un “nuevo volcán” o de “un lejano respiradero de los actuales” (La Nueva Prensa, 1924). Esta supuesta actividad volcánica del Turrubares despertó el interés de José Fidel Tristán, quien dudaba si realmente era posible y a su vez sospechaba que podría tratarse de otra cosa, ya que el notó que los reportes de las personas tenían la particularidad de describir que “el humo se veía salir inmediatamente después que se sentía algún temblor [el subrayado es original]” (Tristán, 1927a, folio 13). El 11 de marzo de 1924 la comisión científica (aún sin Karl Sapper), informó al Ministro de Fomento lo siguiente: 1) que en el cerro Turrubares no había manifestaciones volcánicas, sino numerosos derrumbes que levantaban nubes de polvo y 2) que la zona epicentral del terremoto del 4 de marzo se encontraba al sur de Orotina y visiblemente tenía un origen tectónico (Diario de Costa Rica, 1924d; Gobierno de la República de Costa Rica, 1924; Tristán, 1927a, folios 27, 28;).

En el siglo XIX el cerro Turrubares no solo fue considerado como un volcán, en tanto es un cerro prominente de 1756 m de altitud en medio de muchos otros cerros más bajos, sino que también se le conoció con otros nombres. Stephens lo llamó volcán de San Pablo en 1840 (Fernández, 2002, p. 37) al igual que Marr en 1853 (Marr, 2004, p. 316) y Meagher en 1858 (Fernández, 2002, p. 279). Marr en 1853 (Marr, 2004, p. 316) y Meagher en 1858 lo distinguió del mítico volcán Herradura (Fernández, 2002, p. 279). Montessus de Ballore (1884, p. 208) también se refirió al Turrubares como volcán, pero Pittier (1890) lo corrigió y negó que fuera un volcán (Mora, 2024). El geólogo William Gabb fue categórico en indicar que el pico de Herradura, refiriéndose al cerro Turrubares, no era un volcán (Denyer y Soto, 1999). La idea de que el Turrubares era un volcán prevaleció a inicios del siglo XX (Noriega, 1904, p. 108, p. 224), pero en la segunda mitad de ese siglo se comprobó, geológicamente, que no es un volcán y se cartografió como rocas volcánicas antiguas en diferentes mapas (Denyer y Arias, 1990; Dóndoli *et al.*, 1968; Gobierno de Costa Rica, Dirección de Geología, Minas y Petróleo, 1963; Sandoval *et al.*, 1982).

5.2. Segundo periodo: 1925 a 1932

Un reporte del periódico *La Tribuna* y otro de José Fidel Tristán coinciden sobre una gran erupción del volcán Irazú el 8 de junio de 1925 en la mañana, pero que no pudo ser observada en San José por estar nublado; además habría generado caída de cenizas cerca del volcán según informó el telegrafista de Tierra Blanca (Tabla 5 del [Apéndice A](#), Figura 1c). Estos reportes, aunque coinciden, son aislados, ya que otros rotativos no informaron al respecto y, más aún, tampoco hay reportes de caída de ceniza desde otras localidades, lo que sería de esperar si la erupción hubiese sido grande y realmente hubiera habido dispersión de ceniza hacia sus faldas.

En agosto e inicios de septiembre de 1925, las erupciones en el Irazú se describen como pequeñas y acompañadas de un ruido sordo, según una carta de José Fernández a José Fidel Tristán; sin embargo, estudiantes del Liceo de Costa Rica que ascendieron al volcán el 6 de septiembre de 1925 observaron más bien columnas de vapor que emergían del cráter M tras ese ruido (Tabla 5 del [Apéndice A](#)).

El 24 de septiembre de 1925 *La Tribuna* que informó de una gran erupción que habría durado dos horas, pero fue desmentido por José Fidel Tristán, quien además critica acremente al cronista que redactó la noticia (Tabla 5 del [Apéndice A](#)). De hecho, José Fidel Tristán resume que durante 1925 la actividad en el Irazú había disminuido mucho y no había nada anormal, solo algunas “erupciones pequeñas”, razón por la cual no visitaba el volcán (Tabla 5 del [Apéndice A](#)).

A fines de febrero de 1926, José Fidel Tristán, con base en unas fotos que tomó de Edwin Góngora durante una excursión al Irazú, constató que la fuerte erosión estaba ensanchando la parte superior del cráter M (Figura 4f) y dándole forma de embudo con una chimenea algo cilíndrica (Tabla 5 del [Apéndice A](#)). Además, notó que del lado del cráter M no había ceniza donde usualmente caía durante las erupciones, lo que le sugería que no había habido actividad eruptiva en mucho tiempo, aunado a que solo salía “humo blanco” en pocas cantidades y que el olor a gases sulfurosos era escaso (Tabla 5 del [Apéndice A](#)).

El ensanchamiento del cráter M y la salida de humo constante también fue reportado en diciembre de 1927 por el joven excursionista E. van der Laat (La E. corresponde al nombre de Enrique, L. van der Laat com. pers., 2026), alumno de V año del Liceo de Costa Rica, que envió un relato y fotografías a José Fidel Tristán (Tabla 5 del [Apéndice A](#); Figura 4a). También vio, lejos del cráter, arbustos llenos de lodo fresco a lo cual se añade que los peones de la finca Volio informaron que en los últimos meses de 1927 habían ocurrido algunas erupciones grandes acompañadas de retumbos. Sin embargo, no existen noticias en la prensa u otros reportes de comunidades cercanas que consta-

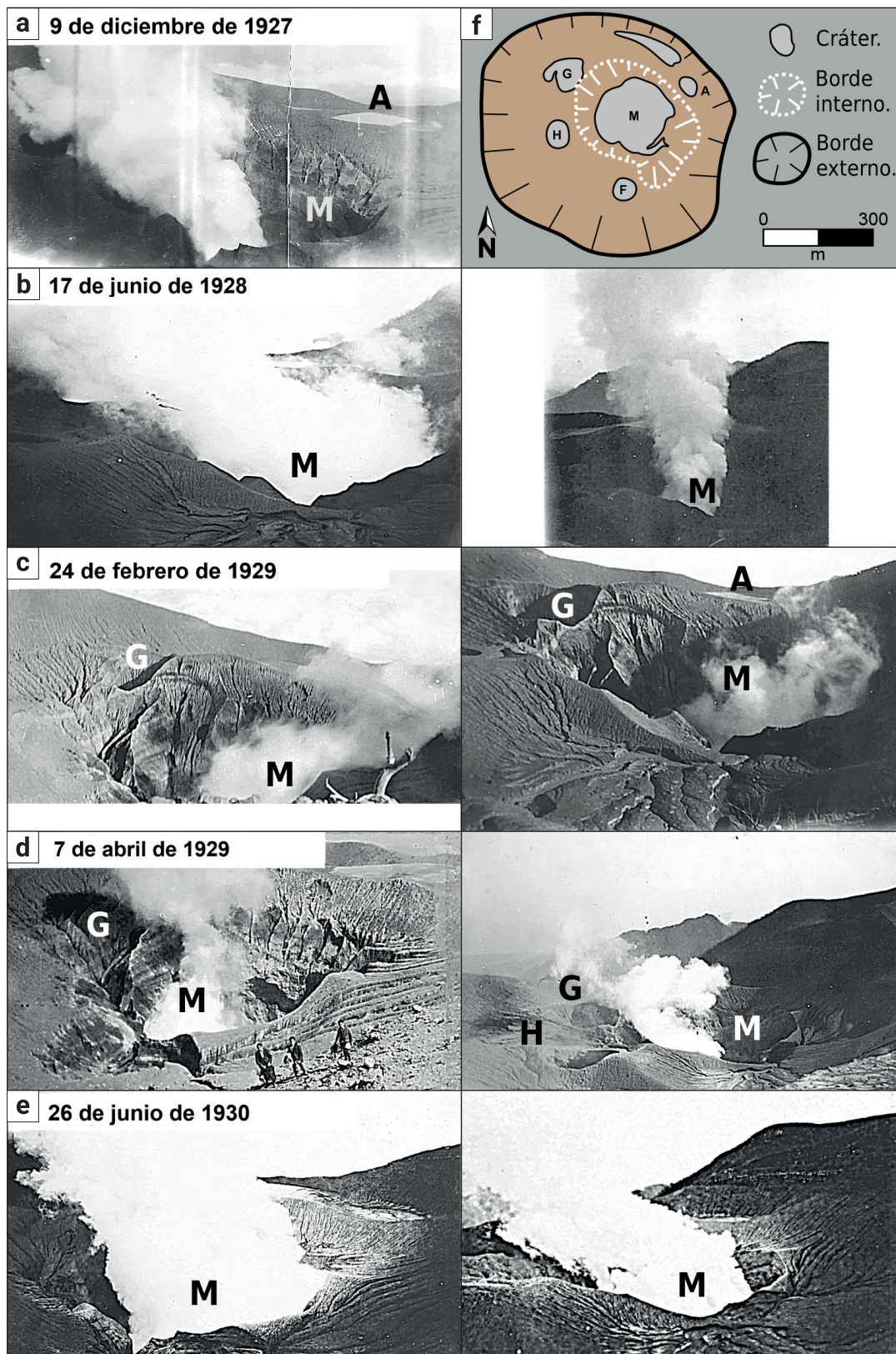


Figura 4. Fotografías del área de los cráteres del volcán Irazú entre 1927 y 1930. a) van der Laet (1927), b) [Dos fotografías del área de los cráteres del Volcán Irazú] (1928), c) Izquierda: Calvo (1929), derecha: Bolaños (1929), d) Izquierda y derecha: [Dos fotografías del área de los cráteres del Volcán Irazú] (1929), e) Izquierda y derecha: Valerio (1930). En cada fotografía se indica el cráter correspondiente según el esquema basado en Fernández (1920) y modificado de Mora *et al.* (2024) que se muestra en f) el cual indica la posición de los cráteres en la cima del volcán Irazú en 1920 como referencia. La nomenclatura de los cráteres de A al G fue dada por Sapper (1901, modificado de la figura 8 de Mora *et al.*, 2024).

ten que hubieran ocurrido erupciones de gran magnitud en esa época. El mismo José Fidel Tristán señala que la actividad no había aumentado mucho en los últimos meses de 1927 (Tabla 5 del [Apéndice A](#)).

El 18 diciembre de 1927 se publicaron unos comentarios de Abelardo Quesada Chacón quien aseguraba que el Irazú se encontraba en “erupción constante”, lo cual se apreciaba siempre en horas de la madrugada y que incluso habían “lluvias de ceniza en la capital” que afectaba plantas que él cultivaba (Tabla 5 del [Apéndice A](#)). Esto contrasta con las observaciones de José Fidel Tristán que daban cuenta que la mayor parte del tiempo hubo actividad débil en el Irazú, generalmente descrita como “humo flojo” o “muy transparente” o “muy ralo” y ocasionalmente hacía referencia a alguna erupción “grande” como sucedió el 3, 17, 19 y 28 de diciembre (Tristán, 1927a, folios 351-366).

Algunos esquemas que dibujó José Fidel Tristán entre 1924 y 1927 muestran columnas que habrían superado fácilmente los 1000 m sobre el nivel del cráter (Figura 5b, c, e, f, g, i, k, m, n, o) o que tienen un patrón de puntos que parece sugerir que contenían ceniza (Figura 5d, i, j). De modo que sería de esperar que por su magnitud, se hubiera reportado alguna caída de ceniza y con ello un mayor impacto mediático pero no fue así. El 25 de abril de 1927, al amanecer, describió una “erupción” como una nube negra y resalta en el esquema la luz del sol detrás de la pluma, lo cual genera que, independientemente si es una erupción o una columna de vapor de agua, esta se vea oscura, de modo que puede generar confusión (Figura 5n). Algo similar ocurrió el 28 de abril de 1927 a juzgar por los colores que utiliza (Figura 5o). Finalmente, José Fidel Tristán describió el 18 de abril de 1925 y el 24 de febrero de 1927 supuestas erupciones cuyas columnas se habrían mantenido poco más de una hora, lo que parece contradecir lo que había criticado al periodista el 24 de septiembre de 1925 (Figura 5e, l). Solo en algunos casos José Fidel Tristán claramente describió columnas de humo no volcánicas, producto de carboneras (p. ej. el 13 de enero 1924, Tristán, 1927b, folio 13) y quemas (p. ej. el 2 de febrero de 1924 y el 8 de marzo de 1926, Figura 5a, h). Lo que se desprende con seguridad de las notas que José Fidel Tristán escribió entre 1925 y 1927 es que el Irazú mantuvo una actividad exhalativa débil e intermitente, que ocasionalmente se tornaba más vigorosa.

En febrero de 1928 una publicación del *Diario de Costa Rica* informó sobre una erupción con una “poderosa columna de humo negro” (Tabla 5 del [Apéndice A](#)). Empero, se indica que esto fue observado a las 4:30 a. m., cuando todavía está oscuro, ya que en febrero generalmente el sol sale entre las 5:30 y 6:00 a. m. en Costa Rica, de modo que es posible confundir una columna de vapor de agua y gases con una de ceniza; además no se encontró información al respecto en otros periódicos que sustenten que hubiera ocurrido una erupción. De hecho, el 21

de mayo de 1928 el mismo rotativo informó sobre “continuas erupciones que se elevan a gran altura” pero en la misma noticia describe que son “bastante blancas” y llegaban a confundirse con las nubes (Tabla 5 del [Apéndice A](#)). También, y sin mayor detalle, José Fidel Tristán observó una columna vertical sobre el Irazú a las 5:20 a. m. del 21 de mayo de 1928 y la describe simplemente como un “espectáculo” lo cual, muy posiblemente, se trataba de una columna de vapor de agua y gases, tal y como se observa en fotografías de ese año (Tabla 5 del [Apéndice A](#); Figura 4b).

La actividad exhalativa se mantuvo entre 1929 y 1931, como lo revelan las noticias que describen una “actividad constante” y emisión de “humo blanco” que era catalogada erróneamente como erupciones (Tabla 5 del [Apéndice A](#); Figura 4c, d y e). En noviembre de 1929, publicó una foto de la actividad del volcán Irazú en la que se observa la pluma siendo dispersada posiblemente hacia el noroeste o norte, si se considera que la fotografía se habría tomado desde San José (con base en la morfología del volcán) pero no se reporta caída de ceniza (Figura 6). Además, en marzo de 1930 se reportó que caía ceniza en el cerro Cabeza de Vaca (Figura 1c), en la finca de Alfredo Volio, pero, a unos 5 km al oeste del cráter activo del Irazú; sin embargo, es un reporte aislado y no se puede confirmar (Tabla 5 del [Apéndice A](#)). Por otra parte, las crónicas de excursionistas, algunos de ellos académicos, que visitaron la cima del volcán en 1929 y 1930, describen que había una erosión intensa de los cráteres (que ya se reportaba en años anteriores) y una profusa actividad hidrotermal caracterizada por: ruidos continuos, trepidaciones y retumbos, sonidos de oleaje, posiblemente erupciones freáticas que lanzaban sedimento y cenizas hacia las paredes del cráter, lo que implica la presencia de un lago activo (Tablas 5 y 6 del [Apéndice A](#), figura 4c, d y e).

En noviembre de 1932 se publicó una noticia que reportaba que la ceniza estaba afectando el ganado en Coronado (*Diario de Costa Rica*, 1932). Sin embargo, no se tienen reportes de actividad volcánica del Irazú en ese año. Una posibilidad es que se tratara de ceniza re trabajada arrastrada por el viento, lo cual es posible ver en época seca y los terrenos están desprovistos de cultivos, aunque no es común que en Coronado haya alguna afectación por ese tipo de nubes de polvo.

5.3. Tercer periodo: fase eruptiva de 1933

En junio de 1933 hubo un aumento en la actividad del volcán Irazú y se reportó caída de ceniza cerca del cráter activo y en San José. Se tiene certeza de que hubo erupciones entre el 12 y el 22 de junio y, si bien atrajo la atención, las descripciones son escasas y hace difícil una interpretación robusta sobre su naturaleza y dinámica (Tabla 7 del [Apéndice A](#)). Los reportes describen que

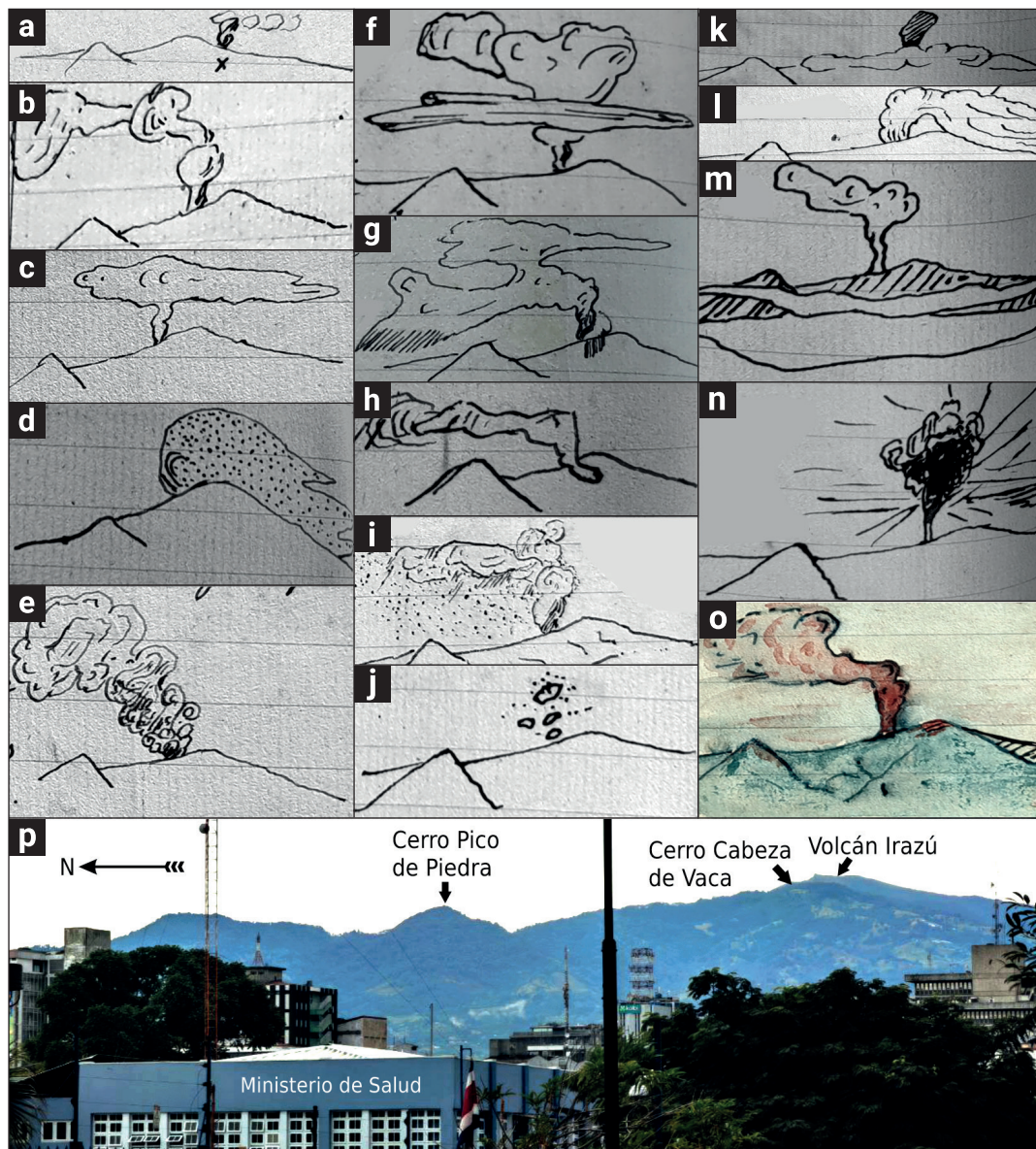


Figura 5. Esquemas de la actividad del Irazú elaborados por José Fidel Tristán entre 1924 y 1927. a) 2 de febrero de 1924 (Tristán, 1927b, folio 33), las equis señalan el sitio donde observó una gran quema. b) 1° de mayo de 1924 (Tristán, 1927b, folio 122), a las 6:30 horas “seguramente varias erupciones por la forma de la columna”. c) 13 de enero de 1925 (Tristán, 1927b, folio 13), menciona que a las 6:40 horas estaba despejado y sospecha que antes debió haber una erupción grande. d) 24 de enero de 1925 (Tristán, 1927b, folio 24), la describe como una nube amarillenta con “todo el aspecto de contener ceniza”. e) 18 de abril de 1925 (Tristán, 1927b, folio 109), a las 6:40 horas vio “salir una erupción muy grande que formó un enorme penacho arremolinado” y así se mantuvo hasta las 8:10 horas cuando empezó a nublarse. f) 2 de junio de 1925 (Tristán, 1927b, folio 154), a las 6:20 horas indica que había una erupción grande, tapada en parte por una nube. g) 8 de junio de 1925 (Tristán, 1927b, folio 160), a las 6:30 horas describe una “grandísima erupción”. h) 8 de marzo de 1926 (Tristán, 1927b, folio 68), a las 8:00 horas notó una columna de humo que parecía una erupción y alarmó a varias personas, pero resultó ser una quema. i) 24 de agosto de 1926 (Tristán, 1927b, folio 237), a las 6:00 horas describe un “gran velo de color amarillo producido por polvo fino” debajo de la columna. j) 6 de febrero de 1927 (Tristán, 1927b, folio 37), describe erupciones pequeñas muy divididas y que dejan una delgada capa de ceniza. k) 21 de febrero de 1927 (Tristán, 1927b, folio 52), a las 16:50 horas describe un cordón de nubes blancas “y se vio salir una erupción de humo oscuro” que duró pocos minutos. l) 24 de febrero de 1927 (Tristán, 1927b, folio 55), a las 6:50 horas vio una “enorme erupción” que doblaba hacia el sur y continuó por media hora y a las 8:10 horas una nube tapó la erupción. m) 20 de marzo de 1927 (Tristán, 1927b, folio 80), a las 17:30 horas observó “una erupción bastante grande”. n) 25 de abril de 1927 (Tristán, 1927b, folio 116), a las 6:30 horas describe una erupción grande y densa cuya parte central se veía negra. o) 28 de abril de 1927 (Tristán, 1927b, folio 119), describe una columna blanca y en la tarde generaba un “lindo espectáculo” con la luz de la puesta del sol. El cerro prominente de forma triangular que Tristán siempre dibuja a la izquierda de los esquemas corresponde al cerro Pico de Piedra. p) Fotografía actual para comparación tomada por Giovanni Peraldo desde San José, cerca del Ministerio de Salud (edificio color celeste) en la cual se aprecia el cerro Pico de Piedra, el cerro Cabeza de Vaca y el volcán Irazú (ver figura 1c).



Figura 6. Fotografía tomada el 13 de noviembre de 1929 (ABC, 1929). Nótese el cerro Pico de Piedra (ver Figuras 1c y 5p), primer cerro prominente de forma triangular a la izquierda de la pluma, mismo que José Fidel Tristán dibujaba en sus esquemas.

las columnas eruptivas pudieron alcanzar entre 400 m y 2000 m, aunque es posible que este último dato esté sobreestimado, ya que de las pocas fotografías que existen se puede notar que la columna es rápidamente dispersada por el viento a una altura de pocos cientos de metros (Tabla 7 del [Apéndice A](#); Figura 7).

La ceniza que llegó a San José fue poca y generalmente fina, lo que denota que la actividad eruptiva era débil. El 17 de junio se informó que se habían recolectado, en San José, entre 63 y 64 g/m² de ceniza equivalentes a 3 mm de espesor en un día, lo cual, como veremos es exagerado (Tabla 7 del [Apéndice A](#)). En el *Diario de Costa Rica* se informó de 1250 libras por hectárea, así como de 2 oz de ceniza caída en 24 h, lo que equivale en ambos casos a unos 57 g/m² y es consistente con el dato anterior, en tanto que otra noticia de *La Tribuna* informó sobre 36 g/m² recolectados en el Centro Nacional de Agricultura en San José (Tabla 7 del [Apéndice A](#); Chaves, 1933). Si se considera una densidad de 1,5 g/cm³, típica de una ceniza de caída vesicular y porosa (la densidad de una andesita ácida es de ~2,7 g/cm³ y una porosidad típica de un depósito de caída distal es de 45 %, lo que da una densidad aproximada de 1,5 g/cm³) el espesor real para los 63 a 64 g/m² de ceniza sería de 0,04 mm y no de 3 mm como se confirmó anteriormente. Jiménez (1933) y Schaufelberger y Jiménez (1933) estimaron en 3 250 000 toneladas el volumen de ceniza emitida por el Irazú del 16 al 22 de junio de 1933 (si se usa una densidad de 2,7 g/cm³ para roca densa, ese volumen corresponde a 1,2 × 10⁶ m³; o bien ~2,2 × 10⁶ m³ de ceniza porosa). Esa cantidad podría equivaler a un espesor constante de 1 cm de ceniza densa o 1,8 cm de ceniza porosa distribuido en una distancia de 20 km (p. ej. desde el cráter activo hasta el sector más oriental del Valle Central) y un ancho de unos 6 km. No obstante, si se toma el valor de los 0,04 mm por día, el espesor

total serían apenas 0.28 cm entre el 16 y 22 de junio en San José. Rubén Torres Rojas describió 1 cm de espesor sobre las hojas y los troncos en las cercanías del volcán, lo cual contrasta con otra descripción realizada entre el 21 y 22 de junio, que se refiere a un espesor de 2 pulgadas (5 cm; Tabla 7 del [Apéndice A](#)) en algunos potreros de Coronado (Figura 1c). Ambos datos están sobreestimados, pero el primero es más coherente con los cálculos anteriores y la disminución del espesor de la ceniza con la distancia. A modo de contraste, para la fase eruptiva de 1963-65 se estimó en 27 kg/m² de ceniza caída en San José (Instituto Costarricense de Electricidad, 1965) lo que equivale, si se usa una densidad de 1500 kg/m³, a 2 cm de espesor.

Profesores y científicos de la época como Carlos Borel, Rubén Torres y el geólogo Paul Schaufelberger visitaron los cráteres del Irazú y coinciden en que no observaron balísticos, lo que es consistente con la opinión de Arturo Volio, dueño de una de las fincas cercanas al cráter activo, quien desmintió las noticias que hablan de “caída de piedras” y que se habían visto “llamas” (Tablas 7 y 8 del [Apéndice A](#)). Rubén Torres y Paul Schaufelberger, observaron derrumbes en las paredes internas del cráter activo, lo que los condujo a interpretar que la ceniza emanada pudo ser generada por el colapso del material y no por la actividad eruptiva (Tabla 8 del [Apéndice A](#)).

Durante la fase eruptiva de 1933 se publicaron algunas descripciones macroscópicas de las cenizas. El Ing. Rafael Tristán describió en *La Tribuna* las cenizas de las primeras erupciones como “arenas negruzcas” en tanto que aquellas de erupciones posteriores eran “finamente pulverizadas como el cemento”, también resalta diferentes colores en la ceniza: “unas veces son negruzcas, otras grises y otras rojas como el polvo de ladrillo” (Tabla 7 del [Apéndice A](#)). El *Diario de Costa Rica* comenta

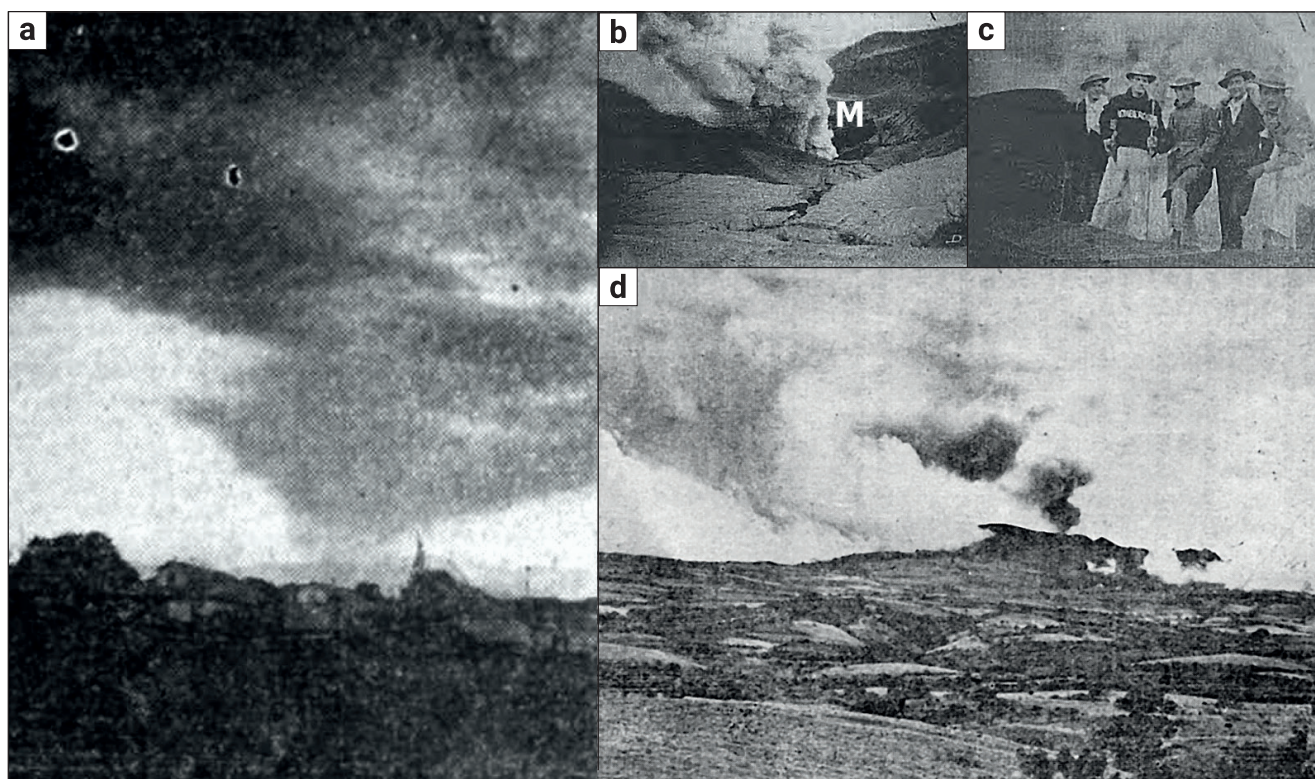


Figura 7. Erupciones de junio de 1933. a) Foto de la erupción del 16 de junio de 1933 a las 4:45 a. m. tomada por Paco Hernández desde el barrio González Lahmann en San José (La Tribuna, 1933a, 1933b, 1933c). Fotos tomadas el 18 de junio de 1933 (Diario de Costa Rica, 1933a), en b) se indica el cráter M o principal según la nomenclatura de Fernández (1920); salen en c) Francisco Campos J., Alberto Morúa, Marco T. Carazo, José Rafael Morúa y Bolívar Alfaro. d) Foto publicada en Schaufelberger y Jiménez (1933) y La Tribuna (1933b).

que las cenizas contienen “partículas blancas y grises” y Rubén Torres Rojas describió acumulaciones de ceniza en la vegetación en la cima del volcán e indica que la primera que cayó era de color “amarillento rojizo” y el superficial gris oscuro (Tabla 7 del [Apéndice A](#)).

Algunos estudios sobre la composición de la ceniza llevaron a un debate científico interesante. En *La Tribuna* publicó, sin indicar la fuente de la información, que la ceniza tenía un 80 % de cuarzo (un valor muy alto que podría darse por una confusión entre la plagioclasa fresca y el cuarzo) y, en la misma noticia, se refieren a una entrevista al Ing. Rafael Tristán donde habla de sus análisis de la ceniza e indica la existencia de hierro metálico en pequeñas cantidades y muy poco o nula cantidad de azufre lo que lo lleva a concluir que las cenizas son neutras y compuestas por silicatos y cuarzo (Tabla 7 del [Apéndice A](#)). Por su parte, el químico del Centro Nacional de Agricultura puso en evidencia el error de Rafael Tristán de confundir el hierro metálico con el óxido ferroso-férrico o magnetita y además reporta que la ceniza contiene un 38,3 % de sílice (Chaves, 1933). Finalmente, otro análisis reporta un 58,52 % de sílice en la ceniza (Diario de Costa Rica, 1933b). Este último dato parece más acertado con la química típica de las cenizas del Irazú, que normalmente

fluctúan entre los 53 y 60 % de sílice (Murata *et al.*, 1966; Alvarado *et al.*, 2006).

5.4. Cuarto periodo: fase eruptiva de 1939 y 1940

Después de la actividad de junio de 1933, el volcán Irazú se mantuvo en calma eruptiva o, al menos, con una actividad poco perceptible restringida al cráter activo (Figura 8). Fernando Torres Vincenzi realizó excursiones a los cráteres del volcán en febrero, agosto y noviembre de 1937, durante las cuales constató: (1) emanaciones de vapor de agua y gases en el cráter G; (2) la presencia de un lago activo en el cráter M, que no solo “parecía hervir”, sino que además presentaba grandes burbujas de gas y producía un ruido característico de ebullición; y (3) la salida de agua desde las paredes del cráter que alimentaba dicho lago (Tabla 9 del [Apéndice A](#)).

El 28 de enero de 1939, Fernando Torres Vincenzi visitó nuevamente los cráteres del Irazú (Figura 8b, c y d) y observó varios cambios que lo llevaron a concluir que la actividad del volcán se había incrementado. En particular, describió: (1) una emanación más profusa de vapor de agua y gases en los cráteres G y M, incluso de forma simultánea, lo que lo llevó a sugerir una

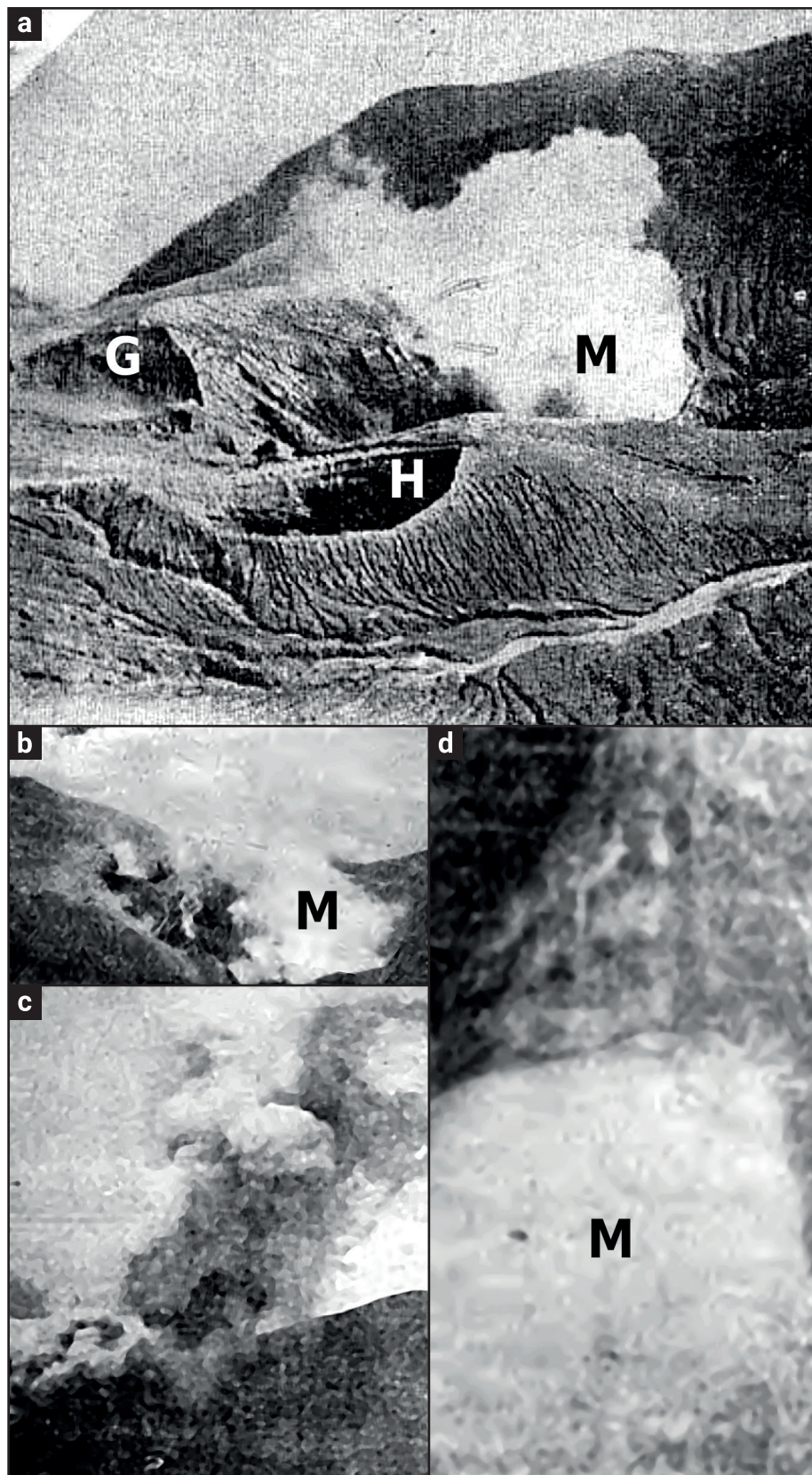


Figura 8. a) Foto tomada el domingo 15 de enero de 1939 en la que se observa la actividad exhalativa (Novedades, 1939a). b, c y d) Fotografías publicadas el 5 de febrero de 1939 en las que se observa la actividad y la laguna en el fondo del cráter M, posiblemente. Aunque la nota la escribió Fernando Torres Vincenzi, la publicación no refiere la autoría de las fotos (Tabla 9 del [Apéndice A](#)). En cada foto se indica el cráter según la nomenclatura de Sapper (1901) y Fernández (1920).

posible interconexión entre ambos (quizás mediante la extensión del lago en el fondo de los cráteres); (2) un aumento de la actividad en el lago, que “parecía hervir con más fuerza y se formaba enorme oleaje”, lo cual atribuyó a un incremento en la temperatura; (3) el depósito de “grandes masas” de azufre en las paredes del cráter; y (4) la presencia de manchas amarillentas, interpretadas como “posiblemente azufre flotante” (Tabla 9 del [Apéndice A](#)).

Un relato de Rubén Torres Rojas publicado el 29 de octubre de 1939 en *La Tribuna* señala que ya a finales de junio de 1939 el volcán mostraba señales de reactivación: la laguna del fondo del cráter era claramente visible y sus aguas se encontraban en movimiento continuo, con aspecto “hirviente”, aunque sin los ruidos fuertes asociados a erupciones (Tabla 9 del [Apéndice A](#)). Por otro lado, para José Merino y Coronado y César Dóndoli, científicos de la época, la actividad en el Irazú en junio de 1939 no parecía tan intensa, pese a reconocer que era mayor que en años anteriores, y solo se observaba la emanación de una columna con una temperatura de alrededor de los 100°C, compuesta de vapor de agua, gases azufrados (“hidrógeno sulfurado, anhídrido sulfuroso y sulfatos (ácido sulfúrico)”, H_2S , SO_4 y H_2SO_4 , respectivamente) cuyo olor se sentía hasta casi una decena de kilómetros del cráter (Tabla 9 del [Apéndice A](#)).

Más de un mes antes de la erupción del 10 de octubre se reportaban caídas de ceniza en los alrededores del volcán, especialmente hacia el suroeste, lo que sugiere, con base en fotografías tomadas el 8 de octubre de 1939, una fase inicial de actividad eruptiva débil a moderada (Figura 9a y b). La erupción del 10 de octubre habría alcanzado casi los 2 km de altura, lo que la hizo muy visible desde San José gracias a las condiciones favorables del tiempo (Figura 9c, d y e). Se reportó caída de ceniza en varias localidades del Valle Central, entre ellas Escazú, Villa Colón (hoy Ciudad Colón) y Puriscal (Figura 10). Durante la visita del 20 de octubre, Rubén Torres Rojas describió una actividad más intensa, caracterizada por emisiones continuas de gases y ceniza, ruidos eruptivos y columnas que se elevaban y colapsaban al disminuir la presión de los gases. En ese momento la emisión gaseosa era tan densa que impedía observar la laguna del fondo del cráter. Asimismo, el mismo observador destacó la aparición de manifestaciones “geiseriformes”, con columnas ocasionales de agua termal semejantes, aunque de menor magnitud, a las observadas en el volcán Poás (Tabla 9 del [Apéndice A](#)).

Luego de un corto periodo de calma entre noviembre y diciembre de 1939, la primera mitad de enero se reportó importante actividad eruptiva en la que las columnas de vapor, gases y cenizas finas se tornaron más conspicuas (Tabla 9 del [Apéndice A](#)). El 18 de enero se reportó caída de ceniza fina en San José y en agosto se constató una importante actividad hidrotermal que generaba columnas de vapor de agua y gases sulfurosos (Tabla 9 del [Apéndice A](#)).

6. Impacto de las fases eruptivas de 1933 y 1939-1940

6.1. Ganadería y producción de leche

Uno de los aspectos que generó más preocupación durante las fases eruptivas de 1933 y de 1939-1940 fue el impacto en la ganadería (Tabla 10 del [Apéndice A](#)), considerando que en las faldas del Irazú había importantes fincas dedicadas principalmente al ganado lechero y que eran propiedad de personas adineradas, como Arturo Volio, Alfredo Volio, Arturo González, Emilio Robert, Bernardo Montes de Oca, Cristóbal Jiménez, Franz de Tattenbach, algunas también pertenecientes, en esa época, a la clase política de Costa Rica.

En ambos periodos eruptivos surgieron noticias en distintos periódicos que informaban que la ceniza caída en los pastos estaba obligando a los finqueros a trasladar el ganado hacia otros terrenos libres de ceniza, a veces a mucha distancia, como Copey ubicado ~40-50 km al sur, o una finca en “Pirrisito”, sin embargo esta localidad no existe, por lo que puede ser un error tipográfico y referirse a Birrisito, ubicado unos 15 km al sur del cráter activo del Irazú (Figura 1c). Esto implicaba una inversión importante aún para los grandes hacendados por los costos de traslado y alquiler de los terrenos, siendo este último incierto porque dependía de la evolución de la actividad volcánica, pero más importante aún, no era una solución aplicable al ganado de raza fina, ya que estaba aclimatado y trasladarlo era igualmente perjudicial, por lo que debía mantenerse encerrado y alimentado con pasto de corta (Tabla 10 del [Apéndice A](#)). Es notable también que las noticias informaron solo sobre el impacto a los grandes hacendados y no se refieren a la suerte que corrían los pequeños y medianos productores, quienes posiblemente no tenían los medios económicos para trasladar su ganado. Solo *La Prensa Libre* publicó un dato que refería que las pérdidas ascendían a más de 100 mil colones entre varios hacendados y señalaba específicamente el caso de Arturo Volio, Alfredo Volio y Cristóbal Jiménez (Tabla 10 del [Apéndice A](#)). Dicha cifra, sin embargo, traída a valor presente equivaldría a muchos miles de millones de colones, de modo que la pérdida reportada por el rotativo podría ser una cifra muy sobreestimada.

Lo que sí es claro es que no hubo muerte de cabezas de ganado durante las fases eruptivas de 1933 y 1939-1940. El 15 de junio de 1933, *La Tribuna* informó que la ceniza había provocado la muerte de cabezas de ganado en las fincas de Emilio Robert y Alfredo Volio, lo cual fue desmentido por este último en una entrevista publicada el 17 de junio por el mismo periódico (Tabla 10 del [Apéndice A](#)). Otra noticia publicada también en *La Tribuna* el 18 de junio de 1933 relata la visita del Dr. A. Rivera a una finca cercana al volcán para investigar una enfermedad que afectaba al ganado; allí describe



Figura 9. a y b) Erupciones del domingo 8 de octubre de 1939 tomadas por Fernando Cañas Vargas y publicadas en Vargas (1967). Vistas desde San José de la erupción del 10 de octubre de 1939 en la mañana publicadas en: c) Vargas (1967), d) Diario de Costa Rica (1939) y e) La Tribuna (1939) tomada desde las terrazas del Palacio de Salubridad Pública (hoy Ministerio de Salud) en San José, ubicado en el centro de la ciudad (coordenadas 9,9320° N y 84,0862° O; comparar con la foto de la Figura 5p).

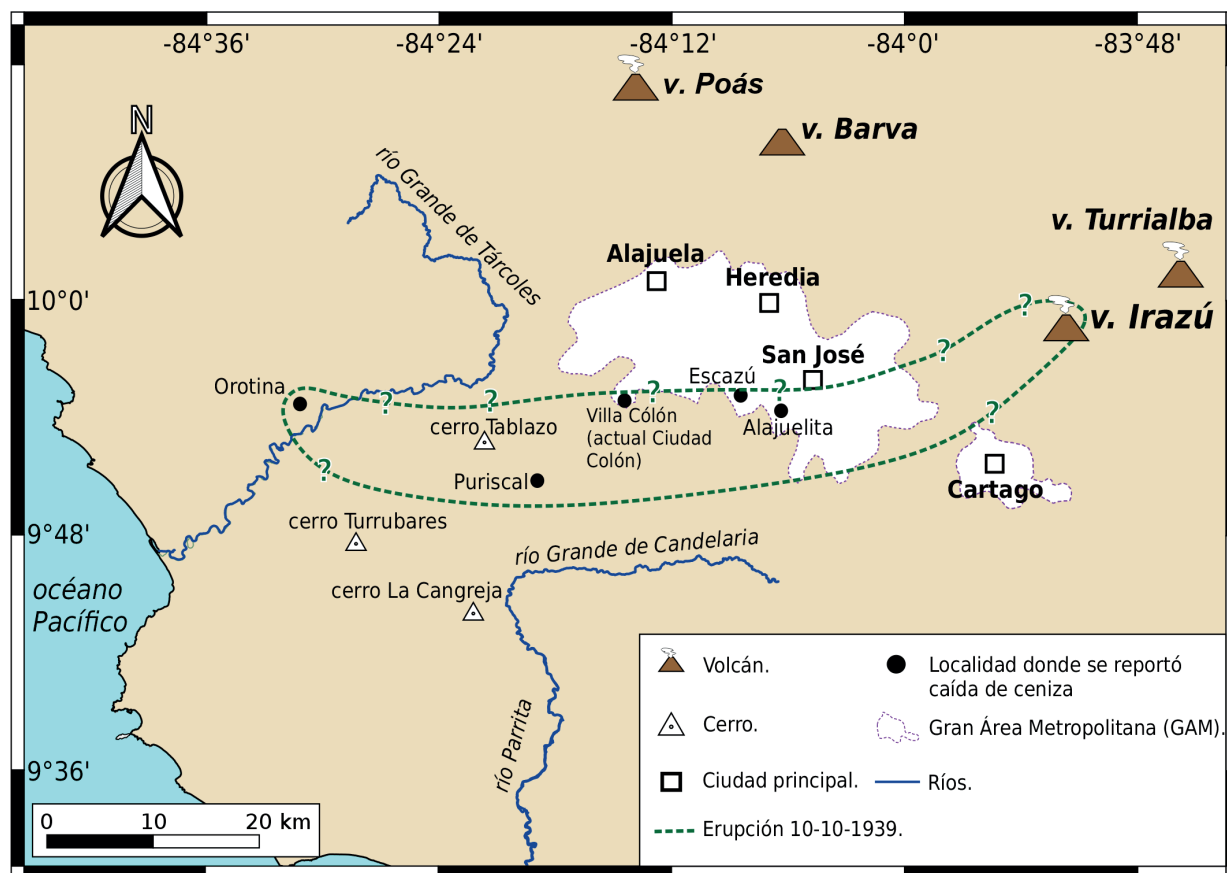


Figura 10. Dispersión de la tefra de caída de la erupción del 10 de octubre de 1939 con base en los reportes publicados en Novedades (1939c).

varios animales enfermos con síntomas como fiebre, trastornos respiratorios y digestivos, y debilidad general, y señala que seleccionó y sacrificó deliberadamente al ejemplar más afectado para practicarle una autopsia in situ, a partir de la cual concluye que la causa era un “proceso lento de intoxicación” provocado por la ingestión de ceniza volcánica, a la que atribuye propiedades cáusticas y tóxicas (Tabla 10 del [Apéndice A](#)). Si bien es correcto que la ceniza puede afectar gravemente al ganado al contaminar pastos y agua, generando un cuadro progresivo de enfermedad, el artículo describe principalmente morbilidad más que mortalidad directa. Una polémica similar ocurrió en noviembre de 1939 por una publicación en el *Diario de Costa Rica* que informaba que el gobierno daba por perdida la ganadería, lo cual fue desmentido categóricamente por Bernardo Montes de Oca, ganadero de la zona y en ese entonces diputado de la República de Costa Rica, quien vehementemente solicitó a la Secretaría de Fomento que rectificara. Al mismo tiempo, Franz de Tattenbach, propietario de la hacienda “Los Guayabillos” en Rancho Redondo (Tabla 10 del [Apéndice A](#); Figura 1c), compartió el mismo criterio, así como también el geólogo César Dóndoli, quien consideraba que el impacto de la actividad eruptiva no era tan severo.

En cuanto a la producción de leche, la información hemerográfica no es suficiente para tener una idea de si realmente la impactó y en qué medida. En junio de 1933 se informó sobre un desmejoramiento en la calidad de la leche que se consumía en San José, debido a que había muchas vacas enfermas y otras que no estaban en condiciones óptimas para producirla; mientras tanto, en 1939 se reportó una disminución en la producción de leche en aproximadamente 4000 botellas y, en consecuencia, un alza en su precio, por lo que se pretendía intensificar la producción, especialmente de las fincas más grandes (Tabla 10 del [Apéndice A](#)).

En ambos periodos el Gobierno de Costa Rica propuso medidas de ayuda al sector. En 1933, el Congreso de Costa Rica aprobó una ley para importar 25 000, colones en forraje y ofrecerlo al sector ganadero bajo condiciones favorables de pago, aunque no es posible saber si la ayuda se materializó. Esta iniciativa fue muy criticada por algunos sectores de corte comunista con el argumento de que favorecía solo a los ganaderos poderosos (Lacteo, 1933; Trabajo, 1933). Por otro lado, en 1940 el gobierno también prometió ayudar al sector ganadero, a lo que la Asociación Nacional de Ganaderos planteó dos medidas de carácter temporal dirigidas a beneficiar únicamente a los ganaderos damnificados con la actividad del Irazú: 1) la exen-

ción temporal de impuestos territoriales; 2) la importación de alimentos concentrados y su venta a precio de costo, excluyendo además el valor del flete ferrocarrilero (Tabla 10 del [Apéndice A](#)). Empero esas solicitudes nunca fueron atendidas (La Tribuna, 1940a;1940b). Queda en duda si esas medidas solo beneficiarían a los ganaderos más poderosos o si lograría permear a aquellos más pequeños. En todo caso, tampoco se pudo constatar si las ayudas se concretaron.

En 1940 surgió una preocupación adicional, el “aborto contagioso” (brucelosis en el ganado) que habría generado pérdidas de consideración en vacas y sementales (Tabla 10 del [Apéndice A](#)). La actividad ganadera de Coronado estaba siendo perjudicada, y aunque aparentemente esta enfermedad no la relacionan con la caída de ceniza, la movilización de ganados de fincas en donde había ganado infectado pudo haber incidido en ello, y habría generado pérdidas en Coronado y, a su vez, un impacto en la producción general de recursos lácteos.

6.2. Agricultura

Desde la colonia, el macizo volcánico ya era usado para la creación de fincas ganaderas y agrícolas. Por ejemplo, se conoce desde el siglo XVII fincas pertenecientes a Jerónimo de Retes, en las alturas de lo que hoy se llama Cerro Retes, y Wilhem Marr en 1852 menciona que el volcán presenta casi hasta la cima dehesas de vecinos de Cartago (Marr, 2004, p. 383). Es así que en la mitad del siglo XX la producción agrícola como papas y otros tubérculos, además de hortalizas desde las dehesas abiertas en el volcán fueron parte de las actividades económicas de la provincia y del país. Cuando ocurrían épocas de erupciones importantes, la actividad se veía afectada, tal como demuestran Mora *et al.* (2024) para el periodo eruptivo de 1917 a 1921. No obstante, la información sobre el impacto de la actividad del Irazú en la agricultura durante las fases eruptivas de 1933 y 1939-1940 es muy escasa. En junio de 1933 se comentaba sobre la posibilidad de pérdidas en diversos productos agrícolas tales como maíz, papas y frijoles, así como también surgió preocupación por la producción de café en las fincas de Tres Ríos, ya que la escasez de lluvias reducía la posibilidad de lavar la ceniza de las hojas de las plantas (Tabla 10 del [Apéndice A](#)). La preocupación por las siembras de café no se volvió a mencionar en las noticias posteriores, muy probablemente porque la erupción fue corta y llegaron algunas lluvias, como lo es usual en junio en el Valle Central, y no es sino hasta 1940 de nuevo surgió producto de la actividad eruptiva (Tabla 10 del [Apéndice A](#)).

De las notas periodísticas recabadas respecto a la agricultura, destaca el hecho de que la mayoría es de tipo especulativo sobre el temor de los agricultores y sobre lo que podría pasar

si la erupción hubiese continuado con vehemencia o por largos periodos, pero no informan de manera explícita sobre pérdidas o áreas específicas afectadas o con destrucción de cultivos (Tabla 10 del [Apéndice A](#)). En contraposición a la posibilidad del impacto en los sembradíos, la ceniza se percibía como algo positivo que a futuro generaría beneficios para la agricultura, tal y como lo señalan Rubén Torres Rojas y César Dóndoli (Tabla 10 del [Apéndice A](#)).

6.3. Salud

Pocas noticias reportan efectos de la actividad eruptiva de 1933 y de 1939-1940 en la salud de la población. En San José se informó que caía polvo fino y como lo hizo continuamente por casi dos semanas, muchas personas se quejaron de molestias en los ojos o en la nariz (Tabla 10 del [Apéndice A](#)). Pero parece haber sido algo muy transitorio y sin efectos duraderos, en tanto no se mencionan más. En octubre de 1939 se hizo mención del efecto de los gases volcánicos en la piel y la respiración (Tabla 10 del [Apéndice A](#)).

También hubo preocupación de parte del Gobierno de Costa Rica por la posible contaminación del agua producto de la ceniza, por lo que se tomó la medida de hacer una inspección, en junio de 1933, con Paul Schaufelberger, geólogo y entonces jefe del Departamento de Geología del Museo Nacional, a partir de la cual se descartó cualquier impacto en el recurso (Tabla 10 del [Apéndice A](#)).

Una noticia del 26 de junio de 1933 en La República, vincula la caída de ceniza intensa con la muerte de varios enfermos de tuberculosis, pero un informe de la Junta de Caridad de San José revela que hubo 163 muertes por tuberculosis en 1933, aunque no se hace ninguna relación explícita con la ceniza del Irazú (Tabla 10 del [Apéndice A](#); Junta de Caridad de San José, 1933, p. 82).

Para las erupciones de junio de 1933 y de 1939-1940, no hay reportes extensos ni prolongados en el tiempo, y solo de manera casi puntual y hasta exagerada en el caso de los supuestos enfermos de tuberculosis fallecidos en 1933 (Tabla 10 del [Apéndice A](#)). Se infiere por tanto, que solo hubo las molestias evidentes y típicas por caída de cenizas y lluvia ácida, pero que no incidieron en efectos duraderos o en la salud pública en general.

6.4. Ecosistema

Un aspecto que difícilmente podría considerarse un impacto es la manera natural que la fauna responde a la actividad volcánica, al trasladarse a áreas no afectadas, sin embargo, en 1933 se trata de plantear esta situación como un problema para la práctica de la caza. Tal vez, se puede pensar que el impacto sería para familias empobrecidas que basaban parte de su dieta a partir de

la “carne de monte” como se le decía al consumo cárnico de especies silvestres. Otro impacto ecológico señalado, esta vez en 1939, es la muerte de aves por las erupciones y la contaminación del ambiente (Tabla 10 del [Apéndice A](#)). Igualmente, en enero de 1940 se mencionó la muerte de animales (Tabla 10 del [Apéndice A](#)).

7. Discusión

Durante las últimas décadas ha sido frecuente citar múltiples erupciones del volcán Irazú entre sus dos principales ciclos eruptivos históricos: 1723–1724 y 1963–1965. No obstante, hasta la fecha no se conocen erupciones durante el siglo XIX, y recientemente se reconstruyó una cronología robusta para el periodo eruptivo de 1917–1921, el único posterior conocido al de 1723–1724 (Mora et al., 2024). En este trabajo se revisó información hemerográfica y documentación de archivos históricos con el objetivo de reconstruir, con la mayor certeza posible, la actividad volcánica del Irazú entre 1924 y 1940.

7.1. ¿Las erupciones de 1924 en el volcán Irazú ocurrieron realmente?

El terremoto de Orotina disparó una serie de noticias contradictorias en la prensa escrita, así como entre funcionarios públicos y científicos de la época, sobre si habían ocurrido o no erupciones en el volcán Irazú. Adicionalmente, el cerro Turrubares se constituyó en un elemento distractor, por cuanto atrajo la atención de la población, particularmente de José Fidel Tristán, quien priorizó investigar las supuestas erupciones que allí ocurrían, no solo a petición del gobierno sino también por interés propio.

De modo que lo que emerge de forma consistente de nuestro análisis es que durante 1924 el volcán Irazú presentaba un sistema hidrotermal activo que generaba actividad exhalativa de intensidad variable, la cual en ocasiones era más notable y podía confundirse con erupciones propiamente dichas. De hecho, el que José Fidel Tristán priorizara la investigación del cerro Turrubares sugiere que el Irazú no presentaba una actividad suficientemente conspicua ni erupciones de ceniza de importancia que volvieran a atraer su atención hacia un volcán que conocía muy bien.

Algunos trabajos previos de revisión histórica del comportamiento eruptivo del volcán Irazú han aceptado directamente, y sin discusión, las noticias (y no todas, pues mencionan solo algunas pocas), de modo que los resultados previos condicionan los siguientes, en este caso respecto a las supuestas erupciones del Irazú en 1924 (p. ej., Barquero, 1975, 1976; Alvarado, 1993; Campos-Durán et al., 2024a, 2024b).

7.2. Actividad del Irazú entre 1925 y 1932

Entre 1925 y 1932 hay numerosos reportes de supuestas erupciones del Irazú que, en gran medida, parecen derivarse tanto de la percepción de una relación entre sismos y erupciones como de la interpretación de diversos fenómenos visibles en el volcán. Entre estos se incluyen incrementos en la emisión de vapor de agua y gases asociados a una mayor actividad hidrotermal. Asimismo, ciertas condiciones meteorológicas locales (por ejemplo, vientos débiles) pueden permitir que una pluma de vapor y gases ascienda a gran altura y adquiera una morfología similar a la de una pluma eruptiva. De igual modo, nubes comunes sobre el volcán podían, dependiendo de la perspectiva del observador, asemejarse a una erupción. En conjunto, estas circunstancias pudieron llevar con frecuencia a interpretar tales fenómenos como erupciones propiamente dichas.

No obstante, no puede descartarse que eventualmente hayan ocurrido desde borbollones, erupciones freáticas de baja magnitud que produjeran únicamente columnas de vapor de agua o incluso erupciones ligeramente más explosivas pero confinadas al cráter, cuya manifestación visible se limitara a una columna de vapor elevándose sobre este. Todas estas formas de actividad han sido observadas en el volcán Poás y, en muchos casos, solo pueden caracterizarse con precisión mediante instrumentación moderna y observaciones continuas (p. ej. van der Laat et al., 2026).

Lo anterior sugiere que el Irazú contaba con un sistema hidrotermal activo y probablemente una acumulación de agua, ya sea en forma de pozas o posiblemente como un lago somero, al interior del cráter principal (denominado M, figura 4f). Esta interpretación resulta plausible, ya que Karl Sapper menciona, aunque de manera vaga, la presencia de un lago en el cráter en 1924 (Tabla 4 del [Apéndice A](#)) y entre 1929 y 1930 las descripciones de los excursionistas reforzaban la idea de un lago por el sonido del oleaje que escuchaban (Tablas 5 y 6 del [Apéndice A](#)). En este contexto, una intensa actividad fumarólica subacuática podría generar el “oleaje” descrito, así como favorecer la acumulación de sedimentos derivados de la erosión de las paredes del cráter. Dichos sedimentos, mezclados con agua hidrotermal, habrían producido el lodo que era expulsado mediante borbollos o pequeñas explosiones freáticas similares a las del volcán Poás (p. ej. van der Laat et al., 2026).

7.3. Naturaleza de las fases eruptivas de 1933 y 1939–1940

Las erupciones ocurridas en junio de 1933 y entre septiembre de 1939 y abril de 1940 pueden corroborarse mediante fuentes hemerográficas y científicas consistentes. Sin embargo, las descripciones de la actividad eruptiva son escasas y poco precisas, y las referencias a la ceniza -disponibles únicamente para la fase

de 1933- son aisladas y en gran medida especulativas.

En junio de 1933 la actividad eruptiva parece haber evolucionado desde una fase inicial de carácter freático -probablemente asociada a la emisión de cenizas claras- hacia una actividad freatomagmática caracterizada por pulsos relativamente ricos en materiales juveniles (escorias negruzcas). Esta evolución sugiere una interacción progresiva entre el sistema hidrotermal y magma ascendente. Es posible que también hayan ocurrido algunos eventos vulcanianos discretos, de carácter casi magmático, responsables de la emisión de las cenizas más oscuras. El volumen total de ceniza emitido, aunque no despreciable, parece haber sido relativamente pequeño, del orden de $1.2 \times 10^6 \text{ m}^3$ en equivalente de roca densa (DRE), lo que sugiere un evento eruptivo de magnitud limitada.

En el caso de la fase eruptiva de 1939 no es posible determinar con precisión su naturaleza, aunque probablemente haya sido similar a la de 1933, pero de menor magnitud, con excepción de la erupción del 10 de octubre de 1939. En conjunto, las fases eruptivas de 1933 y 1939-1940 parecen haber sido mucho más discretas, menos energéticas y de menor duración que la fase eruptiva de 1917-1921, por lo que es posible que no hayan dejado evidencias geológicas.

En este sentido, Mora *et al.* (2024) sugieren que los depósitos observados en la cima del Volcán Irazú, previamente atribuidos a las erupciones de 1939-1940 por Alvarado (1993) y Soto y Sjöbohm (2015), podrían corresponder en realidad a las etapas finales de la fase eruptiva de 1917-1921. De ser así, los depósitos generados durante las fases de 1933 y 1939-1940 -al ser más pequeños y de menor duración- habrían sido posteriormente erosionados.

Esta cuestión resulta difícil de resolver debido a la escasez de afloramientos accesibles en la zona cuspidal del volcán. Asimismo, durante las fases eruptivas de 1933 y 1939-1940 se registraron pocos eventos de caída de ceniza en el área josefina y sus alrededores, y estas no alcanzaron la costa pacífica, como sí ocurrió en la fase eruptiva de 1917-1921. En consecuencia, no es posible reconstruir con certeza las áreas de dispersión de ceniza. Con base en la información disponible, se estima que estas erupciones alcanzaron, a lo sumo, un índice de explosividad volcánica (VEI) de 1 a 2.

7.4. Actividad precursora de las fases eruptivas de 1933 y 1939-1940

En la información analizada existen pocos detalles que arrojen luz sobre la condición del sistema hidrotermal previo a las erupciones de 1933. En cambio, las fuentes disponibles son más descriptivas respecto al periodo previo a las erupciones de 1939. Las excursiones realizadas por Fernando Torres Vincenzi en 1937 a los cráteres del volcán Irazú documentaron emana-

ciones de vapor y gases en el cráter G, así como la presencia de un lago activo en el cráter M, caracterizado por burbujeo vigoroso, liberación continua de gas y aporte de agua desde las paredes del cráter. Este comportamiento es comparable con el observado frecuentemente en la laguna ácida del volcán Poás, donde un sistema freático activo, alimentado por calor y gases desde el fondo, mantiene el lago en continua convección y produce borbollones de agua y lodo ricos en gases (p. ej. van der Laat *et al.*, 2026).

En enero de 1939, el mismo observador registró cambios que interpretó como un incremento en la actividad: emisiones más abundantes de vapor y gases en los cráteres G y M, mayor agitación del lago -con burbujeo intenso y oleaje-, abundante depositación de azufre en las paredes del cráter y la presencia de manchas amarillentas interpretadas como posible azufre flotante. Posteriormente, en junio de 1939, Rubén Torres Rojas también observó mucha agitación en el lago cratérico, posiblemente asociada con procesos convectivos y, posiblemente eventos desde tipo borbollón que podrían generar oleaje, pero no un sonido tan fuerte como el de las erupciones freáticas, lo cual se ha observado en el volcán Poás (p. ej. Rouwet *et al.*, 2014; van der Laat *et al.*, 2026). En conjunto, estas observaciones sugieren un sistema hidrotermal activo que mostró señales de intensificación en los años previos a la fase eruptiva de 1939-1940. Destaca particularmente el incremento en la actividad del lago, evidenciado por su mayor agitación, la probable elevación de la temperatura, la actividad fumarólica subacuática y la presencia de azufre nativo flotante, rasgos muy similares a los observados en la laguna ácida del Poás durante varias fases de actividad en las décadas de 1980 y 1990 y a inicios del siglo XXI (p. ej. Rouwet *et al.*, 2014; Mora Amador *et al.*, 2019).

7.5. Análisis del impacto de las fases eruptivas de 1933 y 1939-1940

El impacto de las fases eruptivas de 1933 y 1939-1940 fue considerablemente menor que el de la fase de 1917-1921; sin embargo, resulta difícil precisarlo y dimensionarlo debido a la escasez de reportes sobre daños. En términos generales, cabe suponer que el impacto de la fase eruptiva de junio de 1933 fue reducida o incluso casi imperceptible, dado su corta duración (aproximadamente 11 días), aunque se registrara caída de ceniza hasta la capital. Los efectos parecieran haberse concentrado principalmente en el sector occidental y suroccidental del volcán Irazú, especialmente en las áreas más proximales al cráter, donde predominaban las fincas dedicadas a la ganadería lechera. En contraste, las fincas situadas hacia el suroeste y sur del cráter, de vocación más agrícola, habrían experimentado impactos nulos o muy leves.

Por su parte, la fase eruptiva de 1939 parece haberse prolongado durante unos siete meses, con episodios intermitentes de actividad, pero sin eventos particularmente energéticos, con excepción de la erupción visible del 10 de octubre. En consecuencia, es probable que sus efectos tampoco hayan sido extensos ni en el espacio ni en el tiempo.

La actividad de 1939-1940 fue además contemporánea a una seguidilla de sismos, algunos posiblemente de profundidad intermedia en la región central y pacífica del país, lo que añadió interés público a las manifestaciones eruptivas. Aun así, estas nunca alcanzaron la intensidad suficiente para generar una alarma generalizada, ni para producir daños significativos o malestar sostenido en la población. Pese a ello, el volcán captó la atención de los gobiernos de turno y surgieron algunas iniciativas orientadas a mitigar posibles impactos, aunque no es posible determinar con certeza si estas llegaron a materializarse. Asimismo, conviene considerar hasta qué punto las quejas del sector ganadero -uno de los más mencionados en las fuentes- reflejaban un impacto real de la actividad del volcán o si, más bien, respondían también a otros intereses o circunstancias, tal como sugieren algunos comentarios publicados por Lacteo (1933) y Trabajo (1933).

8. Conclusiones

El análisis crítico de las fuentes hemerográficas y científicas disponibles no aporta evidencias concluyentes que confirmen la ocurrencia de erupciones en el volcán Irazú entre marzo y abril de 1924 en relación con el terremoto de Orotina. De manera similar, los escasos reportes de supuesta actividad eruptiva entre 1925 y 1932 parecen explicarse principalmente por factores perceptivos, entre ellos la frecuente asociación entre sismos y erupciones o la confusión entre columnas eruptivas y manifestaciones exhalativas del volcán -particularmente plumas de vapor de agua y gases- e incluso nubes meteorológicas observadas sobre la cima.

En conjunto, la evidencia histórica y geológica sugiere que, entre 1925 y 1932, el volcán Irazú tenía un sistema hidrotermal activo, responsable de emisiones vigorosas de vapor y gases y posiblemente asociado con un lago activo al fondo del cráter. Finalmente, las erupciones registradas en 1933 y entre 1939 y 1940 habrían correspondido a episodios eruptivos menores, probablemente de carácter freático a freatomagmático, de corta duración y magnitud limitada (VEI 1-2), muy inferiores en intensidad y alcance al ciclo eruptivo de 1917-1921.

Esta reinterpretación permite conciliar los registros históricos y la escasez de depósitos volcánicos claramente atribuibles a las erupciones de 1933 y 1939-1940 en la zona cuspidal del volcán y pone de relieve la importancia de analizar críticamente las fuentes históricas al reconstruir ciclos eruptivos recientes.

En términos más generales, los resultados sugieren que el Irazú ha sido menos frecuentemente eruptivo en tiempos históricos -desde el siglo XVIII hasta la erupción de 1963- de lo que se ha planteado en diversos trabajos ampliamente citados. La reconstrucción de la actividad del volcán entre 1924 y 1940 contribuye así a mejorar el conocimiento de su historia eruptiva y proporciona una base más sólida para la evaluación de escenarios de amenaza volcánica y la gestión del riesgo en Costa Rica.

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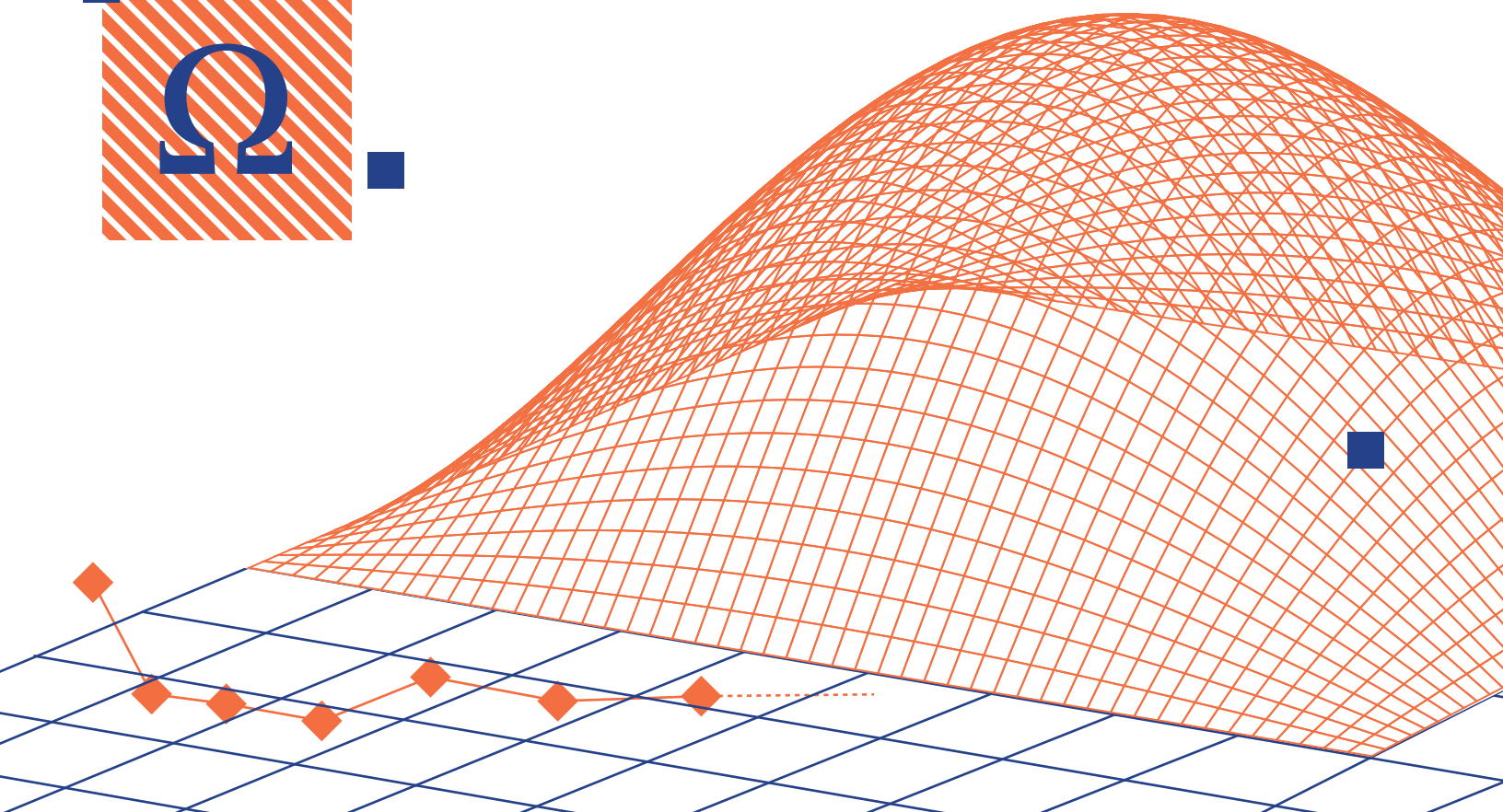
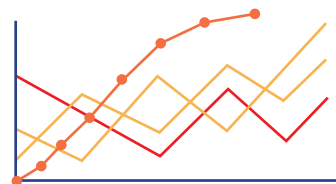
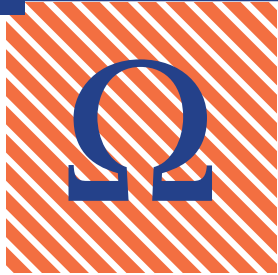
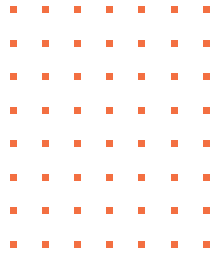


Sección especial

Geofísica Matemática y
Computacional

Special section

Mathematical and Computational
Geophysics



Dynamic Rupture Modeling with a Rate and State Friction Law in a Discontinuous Galerkin Method

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Abstract

Earthquake rupture modeling has proven to be a useful tool to understand the physics behind different slip sequences inferred from observations. However, modeling earthquake dynamics is challenging due to the complex frictional behavior during the rupture evolution. We implement a rate-and-state friction law in DGCrack (Tago *et al.* 2012), a numerical code based on the discontinuous Galerkin method, with a time staggered strategy inspired by the work of (Rojas *et al.* 2009). We propose an ad hoc flux along the fault instead of an upwind flux commonly used in discontinuous Galerkin schemes (e.g. Zhang *et al.* 2023). While this approach generates high-frequency oscillations, we control them through an artificial damping parameter without compromising spectral content. We show that DGCrack correctly reproduces fault waveforms compared with other higher-order state-of-the-art numerical strategies. Our simulations indicate that the artificial damping introduces numerical dispersion and dissipation that are actually smaller than those inherently introduced by an upwind flux.

Key words: Discontinuous Galerkin method; Dynamic rupture; Rate and state friction law.

Resumen

La modelación de la dinámica de la ruptura sísmica ha demostrado ser una herramienta útil para comprender la física detrás de las distintas secuencias de deslizamiento inferidas a partir de observaciones. Sin embargo, modelar la dinámica de los terremotos resulta un desafío debido al complejo comportamiento de la fricción durante la evolución de la ruptura. Implementamos una ley de fricción *rate & state* en DGCrack (Tago *et al.* 2012), un código numérico basado en el método de Galerkin discontinuo, con una estrategia de escalonamiento temporal inspirada en el trabajo de (Rojas *et al.* 2009). Proponemos un flujo ad hoc sobre la falla en lugar del flujo upwind comúnmente utilizado en esquemas de Galerkin discontinuo (e.g. Zhang *et al.* 2023). Si bien este enfoque genera oscilaciones de alta frecuencia, las controlamos con un parámetro de amortiguamiento artificial sin comprometer el contenido espectral. Demostramos que DGCrack reproduce correctamente las formas de onda en la falla en comparación con otras estrategias numéricas de orden superior del estado del arte. Además, mostramos que el amortiguamiento artificial produce una dispersión y disipación numérica en nuestras simulaciones que resultan menores que las introducidas intrínsecamente por un flujo *upwind*.

Palabras clave: Método de Galerkin discontinuo; Dinámica de la ruptura, Ley de fricción "rate and state", Sismología computacional.

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1. Introduction

Understanding the rupture process of earthquakes has been one of the most challenging research topics of the seismological community. Dynamic rupture has been studied from complex experimental setups to computational modeling studies. Among the different friction laws that have been proposed, rate and state friction laws have proven particularly useful for reproducing both short- and long-term behavior of fault dynamics. These friction laws are favored by seismologists to model sequences of earthquakes and aseismic slip, as they can model not only the rupture but the healing process that follows (Erickson *et al.*, 2020; Erickson *et al.*, 2023).

Some of the earliest dynamic rupture numerical simulations were done with a boundary integral equation method (BIEM) (Das and Aki, 1977; Andrews, 1985). This approach has since been continuously developed due to its low computational cost and its suitability for simulating complex geometries (Lapusta *et al.*, 2000; Ando and Kaneko, 2018; Qian *et al.*, 2019; Aochi and Cruz-Atienza, 2025). However, BIEM is limited in its ability to handle heterogeneous media, prompting the development of full-domain discretization strategies. These include the finite difference method (FDM) (Andrews, 1976; Day, 1982; Madariaga *et al.*, 1998; Cruz-Atienza and Virieux, 2004; Dalguer and Day, 2007; Moczo *et al.*, 2007; Rojas *et al.*, 2008; Kozdon *et al.*, 2012, 2013), the finite volume method (FVM) (Benjemaa *et al.*, 2007; Benjemaa *et al.*, 2009), the finite element method (FEM) (Oglesby *et al.*, 1998; Oglesby *et al.*, 2000; Barall, 2009; Galis *et al.*, 2010; Aagaard *et al.*, 2013) and the spectral element method (SEM) (Ampuero, 2002; Kaneko *et al.*, 2008; Galvez *et al.*, 2014).

Among these, the staggered-grid finite difference method (FDM) has been widely adopted due to its computational efficiency and simplicity (Day *et al.*, 2005; Dalguer and Day, 2007). Day and Ely (2002) described an artificial damping parameter for this strategy to suppress spurious high-frequency oscillations (HFOs) due to the abrupt stress drops at the rupture front. This method has been highly successful because of its simple implementation and its ability to preserve the elastodynamic spectrum while selectively attenuating high-frequency oscillations (Day *et al.*, 2005; Rojas *et al.*, 2009).

The discontinuous Galerkin method (DGM) is particularly well suited for earthquake dynamics, as its solutions are inherently discontinuous at element boundaries. DGM combines the high-order accuracy of FEM (Zienkiewicz and Taylor, 2005) with the flux strategies of FVM for handling discontinuous solutions (LeVeque, 2002). The first DGM for elastodynamics, using a linear slip weakening law, was introduced by Benjemaa *et al.* (2007) with constant basis functions in 2D, equivalent to FVM, and later extended to 3D (Benjemaa *et al.*, 2009). De la

Puente *et al.* (2009) developed ADER-DG, a high-order DGM combined with an arbitrarily higher-order derivatives (ADER) time integration scheme (Käser and Dumbser, 2006), for dynamic rupture with higher order approximations in space and time. However, high-order convergence was not achieved for dynamic rupture problems due to the nonlinearity of the fault rupture equations, as also reported by others (Day *et al.*, 2005; Rojas *et al.*, 2008). In ADER-DG, a modal basis is used, and the rupture discontinuity is resolved via an upwind flux (Godunov state). This strategy avoided the appearance of HFOs because of its dissipative nature (Hesthaven and Warburton, 2007). The 3D ADER-DG implementation was later completed by Pelties *et al.* (2012).

Another DGM approach, DGCrack, was simultaneously introduced by Tago *et al.* (2012) as a development of the GeoDG3D parallel code (Etienne *et al.*, 2010). This nodal-basis method uses second order elements next to the fault discontinuity with an ad hoc fault flux. While DGCrack suffers from HFOs, its simplicity yields high computational efficiency and results comparable to those from third-order ADER-DG.

Rate and state friction laws were first coupled with DGM by Pelties *et al.* (2014), who also used an upwind flux to solve the Riemann problem. More recently, Zhang *et al.* (2023) introduced a Mixed-Flux-Based Nodal DGM that also solves the Riemann problem with an upwind flux on fault faces but with central fluxes along non-fault faces to mitigate spatial spike oscillations (SSOs) caused by mesh asymmetry. They avoided central fluxes elsewhere due to potential HFOs, although these may have resulted from strong mesh coarsening. In fact, Etienne *et al.* (2010) demonstrated that dispersion remains controlled when the minimum wavelength resolution requirement, dependent on the approximation order, is satisfied.

In this study, we extend DGCrack to incorporate rate and state friction laws, drawing inspiration from Rojas *et al.* (2009) who proposed a time staggered for the friction coefficient computation. Our aim is to maintain the simplicity of second-order ad hoc fault fluxes while using energy-conserving central fluxes elsewhere. To suppress HFOs, we employ an artificial damping strategy following Day *et al.* (2005). This choice allows us to control energy dissipation, something not achievable with upwind fluxes, while also exploring an alternative approach that remains both simple and computationally efficient. In the following sections, we first introduce the elastodynamic hyperbolic system together with the fault boundary conditions governed by rate-and-state friction. We then describe the discretization of these equations using a DGM, with particular emphasis on the computation of fault fluxes where frictional properties evolve dynamically. The Results section presents the verification of our implementation with the rate-and-state friction benchmark TPV104 from the Statewide California Earthquake Center (SCEC) (Harris *et*

al., 2018). We begin by analyzing the role of artificial damping, followed by waveform comparisons with other state-of-the-art numerical approaches and an assessment of convergence under grid refinement. Finally, we conclude with a discussion of the advantages and limitations of our formulation.

2. Dynamic rupture equations

In dynamic rupture modeling, a friction law is applied along an inner boundary, the fault, embedded within an elastic medium. The following subsections present the governing continuum equations, the associated boundary conditions, and the formulation of the rate-and-state friction law.

2.1. Elastodynamics: First order hyperbolic system

The elastic wave propagation can be modelled with a velocity-stress formulation as a first-order hyperbolic system as:

$$\rho \partial_t \underline{v} = \sum_{\theta \in \{x, y, z\}} \partial_\theta \underline{\underline{M}}_\theta \underline{\sigma}, \quad (1)$$

$$\underline{\underline{\Lambda}} \partial_t \underline{\sigma} = \sum_{\theta \in \{x, y, z\}} \partial_\theta \underline{\underline{N}}_\theta \underline{v}, \quad (2)$$

where ∂_t and ∂_θ are the time and space derivatives for $\theta \in \{x, y, z\}$, ρ is the density, $\underline{\underline{\Lambda}} = \text{diag}(3/(3\lambda + 2\mu), 3/(2\mu), \mu^{-1}, \mu^{-1}, \mu^{-1})$ is the diagonal compliance matrix with the Lamé parameters λ and μ , $\underline{v} = [v_x, v_y, v_z]^T$ is the velocity vector and $\underline{\sigma} = [\omega, \omega', \omega'', \sigma_{xy}, \sigma_{xz}, \sigma_{yz}]^T$ is the stress vector composed by linear combinations of the independent components of the stress tensor, proposed to factorize medium properties (Benjemaa *et al.*, 2009). The first three components of the stress vector are $\omega = \frac{1}{3}(\sigma_{xx} + \sigma_{yy} + \sigma_{zz})$, $\omega' = \frac{1}{3}(2\sigma_{xx} - \sigma_{yy} - \sigma_{zz})$ and $\omega'' = \frac{1}{3}(-\sigma_{xx} + 2\sigma_{yy} - \sigma_{zz})$. The matrices $\underline{\underline{M}}_\theta$ and $\underline{\underline{N}}_\theta$ for all $\theta \in \{x, y, z\}$ are constant real matrices defined in [Appendix A](#).

2.2. Fault boundary conditions and rate and state friction law

Rupture is described by the velocity field discontinuity over a known fault surface, Γ . The fault, Γ , is considered as an inner boundary where dynamic boundary conditions are imposed. It is convenient to arbitrarily identify a plus side (+), and a minus sign (−) of Γ to express the velocity, $\underline{v}$, over Γ as

$$\underline{v}^\pm(t, \underline{x}) = \lim_{\epsilon \rightarrow 0} \underline{v}(t, \underline{x} \pm \epsilon \hat{n}(\underline{x})), \quad (3)$$

where $\epsilon \in \mathbb{R}$, $\hat{n}(\underline{x})$ is the unit normal vector to the fault. For 3D simulations the fault, Γ , is simply a plane, or a set of adjacent planes. In the remainder of the manuscript, we omit the explicit dependence of the fields unless needed for clarity.

The velocity, $\underline{v}^\pm$, can be decomposed in its normal (N) and tangential (T) components as

$$\underline{v}_N^\pm = (\hat{n} \cdot \underline{v}^\pm) \hat{n}, \quad (4)$$

$$\underline{v}_T^\pm = \underline{v}^\pm - \underline{v}_N^\pm = \underline{v}^\pm - (\hat{n} \cdot \underline{v}^\pm) \hat{n}. \quad (5)$$

The discontinuity of the tangential velocity through Γ is known as the slip rate, $\underline{V}$, and is defined as

$$\underline{V}_T := \llbracket \underline{v}_T \rrbracket = \underline{v}_T^+ - \underline{v}_T^-, \quad (6)$$

$$V = \|\underline{V}_T\|, \quad (7)$$

while the discontinuity of the normal component is

$$\underline{V}_N := \llbracket \underline{v}_N \rrbracket = \underline{v}_N^+ - \underline{v}_N^-, \quad (8)$$

$$V_N = \|\underline{V}_N\|, \quad (9)$$

The frictional strength of the fault, τ , is determined with the Amontons-Coulomb model as

$$\tau = f(V, \psi) \sigma_N, \quad (10)$$

where $f(V, \psi)$ is the friction coefficient which is a function of the norm of the slip rate, V eq. (7), and the state variable, ψ , which can be interpreted as the average lifetime of the asperity population (Dieterich and Kilgore, 1996). This last variable allows modelling the fault healing after rupture, which is crucial for earthquake cycle simulations. When the normal component of the traction, $\underline{T}_N$, is compressive throughout the motion, then the normal stress over the fault can be computed as

$$\sigma_N = \|\underline{T}_N\|. \quad (11)$$

Tensile excursions can likewise be accommodated with only minor modifications (Day *et al.*, 2005). With a rate and state friction law, the modulus of the tangential traction is always equal to the shear strength of the fault, i.e. $\|\underline{T}_T\| = \tau$.

For the friction coefficient we use the regularized rate and state friction law

$$f(V, \psi) = a \cdot \text{asinh} \left[\frac{V}{2V_0} \exp \left(\frac{\psi}{a} \right) \right] \quad (12)$$

where a is a frictional parameter and V_0 is the steady state sliding reference velocity. These types of friction law were first introduced by Dieterich (1979) and Ruina (1983) from small-scale laboratory experiments and have been widely used in numerical modeling of earthquake processes (Rice and Ben-Zion, 1996; Lapusta *et al.*, 2000; Rice, 2001). The state variable evolves according to its own ordinary differential equation (ODE) as

$$d_t \psi = -\frac{V}{L} [\psi - \psi_{ss}(V)], \quad (13)$$

where L is a critical slip distance and $\psi_{ss}(V)$ is the state variable at steady state computed as

$$\psi_{ss}(V) = a \cdot \ln \left\{ \frac{2V_0}{V} \sinh \left[\frac{f_{ss}(V)}{a} \right] \right\}, \quad (14)$$

where $f_{ss}(V)$ is the steady state friction

$$f_{ss}(V) = f_w + \frac{f_{LV}(V) - f_w}{[1 + (V/V_w)^8]^{1/8}}, \quad (15)$$

f_w and V_w are friction parameters and $f_{LV}(V)$ is the low-velocity steady state friction coefficient given by

$$f_{LV}(V) = f_0 - (b - a) \ln(V/V_0), \quad (16)$$

where b is another friction parameter. The rupture evolution strongly depends on the friction parameters a and b . Whenever $b > a$, the fault is velocity weakening; otherwise it is velocity strengthening. A velocity weakening region permits ruptures to nucleate and propagate, and therefore models a seismically strong-coupled surface (whereas rupture cannot nucleate in a velocity-strengthening region).

Finally, the last condition is

$$VT_r = \tau \underline{V}, \quad (17)$$

which indicates that the slip rate must follow the same direction as the tangential traction.

3. Discrete equations with a discontinuous Galerkin method

To solve the hyperbolic system, eqs. (1) and (2), the domain Ω is discretized into a conforming mesh of K elements as

$$\Omega \simeq \Omega_h = \bigcup_{i=1}^K D^i, \quad (18)$$

where D^i is the i -th straight-sided tetrahedron of the mesh.

Following Etienne *et al.* (2010), a nodal Discontinuous Galerkin method with central fluxes and an explicit leap-frog time scheme is applied to eqs. (1) and (2) to obtain

$$\begin{aligned} \rho^i \left(\underline{I}_3 \otimes \underline{\mathcal{K}}^i \right) \frac{\underline{v}^{(i,m+\frac{1}{2})} - \underline{v}^{(i,m-\frac{1}{2})}}{\Delta t} = & - \sum_{\theta \in \{x,y,z\}} \left(\underline{M}_\theta \otimes \underline{\mathcal{E}}_\theta^i \right) \underline{\sigma}^{(i,m)} \\ & + \frac{1}{2} \sum_{k \in N^i} \left[\left(\underline{\mathcal{P}}^{ik} \otimes \underline{\mathcal{F}}^{ik} \right) \left(\underline{\sigma}^{(i,m)} + \underline{\sigma}^{(k,m)} \right) \right], \end{aligned} \quad (19)$$

$$\begin{aligned} \left(\underline{\Lambda}^i \otimes \underline{\mathcal{K}}^i \right) \frac{\underline{\sigma}^{(i,m+1)} - \underline{\sigma}^{(i,m)}}{\Delta t} = & - \sum_{\theta \in \{x,y,z\}} \left(\underline{N}_\theta \otimes \underline{\mathcal{E}}_\theta^i \right) \underline{v}^{(i,m+\frac{1}{2})} \\ & + \frac{1}{2} \sum_{k \in N^i} \left[\left(\underline{\mathcal{Q}}^{ik} \otimes \underline{\mathcal{F}}^{ik} \right) \left(\underline{v}^{(i,m+\frac{1}{2})} + \underline{v}^{(k,m+\frac{1}{2})} \right) \right] \end{aligned} \quad (20)$$

where m is the time index, N^i is the neighbor list of the i -tetrahedron, Δt is the time step size, $\underline{I}_3$ is the identity matrix of $\mathbb{R}^3$, $\underline{\mathcal{K}}^i$ is the mass matrix, $\underline{\mathcal{E}}_\theta^i$ for $\theta \in \{x, y, z\}$ are the stiffness matrices, $\underline{\mathcal{F}}^{ik}$ is the flux matrix and $\underline{\mathcal{P}}^{ik}$ and $\underline{\mathcal{Q}}^{ik}$ are defined in [Appendix B](#). If we use second-order elements, we have a second order numerical scheme in space and time. We follow Tago *et al.* (2012) to separate the flux over the fault and to rewrite eqs. (19) and (20) as

$$\begin{aligned} \underline{v}^{(i,m+\frac{1}{2})} = & \underline{v}^{(i,m-\frac{1}{2})} + \frac{\Delta t}{\rho^i} \left(\underline{I}_3 \otimes \underline{\mathcal{K}}^i \right)^{-1} \left[- \sum_{\theta \in \{x,y,z\}} \left(\underline{M}_\theta \otimes \underline{\mathcal{E}}_\theta^i \right) \underline{\sigma}^{(i,m)} \right. \\ & + \frac{1}{2} \sum_{\substack{k \in N^i \\ D^i \cap D^k \not\subseteq \Gamma}} \left[\left(\underline{\mathcal{P}}^{ik} \otimes \underline{\mathcal{F}}^{ik} \right) \left(\underline{\sigma}^{(i,m)} + \underline{\sigma}^{(k,m)} \right) \right] \\ & \left. + \delta^\Gamma \left(\underline{I}_3 \otimes \underline{\mathcal{F}}^{ik} \right) \underline{\mathbb{F}}^{(\Gamma i, m+\frac{1}{2})} \right], \end{aligned} \quad (21)$$

$$\begin{aligned} \underline{\sigma}^{(i,m+1)} = & \underline{\sigma}^{(i,m)} + \frac{\Delta t}{\rho^i} \left(\underline{\Lambda}^i \otimes \underline{\mathcal{K}}^i \right)^{-1} \left[- \sum_{\theta \in \{x,y,z\}} \left(\underline{N}_\theta \otimes \underline{\mathcal{E}}_\theta^i \right) \underline{v}^{(i,m+\frac{1}{2})} \right. \\ & + \frac{1}{2} \sum_{\substack{k \in N^i \\ D^i \cap D^k \not\subseteq \Gamma}} \left[\left(\underline{\mathcal{Q}}^{ik} \otimes \underline{\mathcal{F}}^{ik} \right) \left(\underline{v}^{(i,m+\frac{1}{2})} + \underline{v}^{(k,m+\frac{1}{2})} \right) \right] \\ & \left. + \delta^\Gamma \left(\underline{I}_3 \otimes \underline{\mathcal{F}}^{ik} \right) \underline{\mathbb{G}}^{(\Gamma i, m+1)} \right], \end{aligned} \quad (22)$$

where δ^Γ is a Dirac delta function to identify fault faces and the fault fluxes, $\underline{\mathbb{F}}^{(\Gamma i, m+\frac{1}{2})}$ and $\underline{\mathbb{G}}^{(\Gamma i, m+1)}$ are defined as

$$\underline{\mathbb{F}}^{(\Gamma i, m+\frac{1}{2})} = \underline{T}^{(\Gamma ik, m+\frac{1}{2})} = \underline{T}_N^{(\Gamma ik, m+\frac{1}{2})} + \underline{T}_T^{(\Gamma ik, m+\frac{1}{2})}, \quad (23)$$

$$\underline{\underline{G}}^{(\Gamma i, m+1)} = \underline{\underline{Q}}^{ik} \underline{\underline{v}}^{(i, m+\frac{1}{2})}, \quad (24)$$

and $\underline{\underline{T}}_N^{(\Gamma ik, m+\frac{1}{2})}$ and $\underline{\underline{T}}_T^{(\Gamma ik, m+\frac{1}{2})}$ are the normal and tangential tractions at the fault computed from the i -tetrahedron at time step m , respectively. The fault flux to update the stress vector, $\underline{\underline{G}}^{(\Gamma i, m+1)}$, only considers the velocity vector from the element in which that stress vector is computed.

If the integration of the Galerkin matrices, $\underline{\underline{K}}^i$ and $\underline{\underline{F}}^{ik}$, is done in the standard element (Zienkiewicz and Taylor, 2005), they can be written as

$$\underline{\underline{K}}^i = \underline{\underline{W}}^i \underline{\underline{K}}^e, \quad \underline{\underline{F}}^{ik} = S^{ik} \underline{\underline{F}}^e, \quad (25)$$

where $\underline{\underline{K}}^e$ is the standard mass matrix, $\underline{\underline{F}}^{ik}$ is the standard flux matrix, W^i is the volume of the i -tetrahedron and S^{ik} is the surface shared by the i - and k -tetrahedra.

Using the standard matrices, $\underline{\underline{K}}^e$ and $\underline{\underline{F}}^{ik}$, we can rewrite the update velocity scheme, eq.(21), as

$$\underline{\underline{v}}^{(i, m+\frac{1}{2})} = \underline{\underline{v}}^{(i, m-\frac{1}{2})} + \underline{\underline{R}}^{(i, m)} + \frac{\delta^\Gamma \Delta t S^{ik}}{\rho^i W^i} \left(\underline{\underline{I}}_3 \otimes \underline{\underline{K}}^{e-1} \underline{\underline{F}}^e \right) \left(\underline{\underline{T}}_N^{(\Gamma ik, m+\frac{1}{2})} + \underline{\underline{T}}_T^{(\Gamma ik, m+\frac{1}{2})} \right), \quad (26)$$

where

$$\begin{aligned} \underline{\underline{R}}^{(i, m)} = & \frac{\Delta t}{\rho^i W^i} \left(\underline{\underline{I}}_3 \otimes \underline{\underline{K}}^e \right)^{-1} \left[- \sum_{\theta \in \{x, y, z\}} \left(\underline{\underline{M}}_\theta \otimes \underline{\underline{E}}_\theta^i \right) \underline{\underline{\sigma}}^{(i, m)} \right. \\ & \left. + \frac{1}{2} \sum_{\substack{k \in N^i \\ D^i \cap D^k \neq \emptyset}} \left[\left(\underline{\underline{P}}^{ik} \otimes \underline{\underline{F}}^{ik} \right) \left(\underline{\underline{\sigma}}^{(i, m)} + \underline{\underline{\sigma}}^{(k, m)} \right) \right] \right], \end{aligned} \quad (27)$$

contains the stiffness term and the non-fault fluxes.

The next subsections will explain how to compute the fault tractions, $\underline{\underline{T}}_T^{(\Gamma ik, m+\frac{1}{2})}$ and $\underline{\underline{T}}_N^{(\Gamma ik, m+\frac{1}{2})}$, that fulfill the rate and state friction law and the inner boundary conditions.

3.1. Fault tangential traction, $\underline{\underline{T}}_T^{(\Gamma ik, m+\frac{1}{2})}$, computation

Assuming that the i -tetrahedron is in the plus side of the fault, the slip rate between two adjacent tetrahedra can be computed from eq. (6) and the tangential projection of the velocity scheme, eq. (26), as

$$\begin{aligned} \underline{\underline{V}}^{(i, m+\frac{1}{2})} &= \underline{\underline{v}}_T^{(i, m+\frac{1}{2})} - \underline{\underline{v}}_T^{(k, m+\frac{1}{2})} \\ &= \underline{\underline{V}}^{(i, m-\frac{1}{2})} + \underline{\underline{R}}_T^{(i, m)} - \underline{\underline{R}}_T^{(k, m)} + \frac{\delta^\Gamma \Delta t S^{ik} (\rho^i W^i + \rho^k W^k)}{\rho^i \rho^k W^i W^k} \left(\underline{\underline{I}}_3 \otimes \underline{\underline{K}}^{e-1} \underline{\underline{F}}^e \right) \underline{\underline{T}}_T^{(\Gamma ik, m+\frac{1}{2})} \end{aligned} \quad (28)$$

The tangential fault traction, $\underline{\underline{T}}_T^{(\Gamma ik, m+\frac{1}{2})}$, can be decomposed in terms of the initial tangential traction, $\underline{\underline{T}}_T^{(\Gamma ik, 0)}$ and the total

shear traction due to the friction evolution, $\underline{\underline{T}}_T^{(\Gamma ik, m+\frac{1}{2})}$, as

$$\underline{\underline{T}}_T^{(\Gamma ik, m+\frac{1}{2})} = \underline{\underline{T}}_T^{(\Gamma ik, 0)} - \underline{\underline{T}}_T^{(\Gamma ik, m+\frac{1}{2})} \quad (29)$$

Then, the slip rate update in terms of the total shear traction can be written as

$$\underline{\underline{V}}^{(i, m+\frac{1}{2})} = \underline{\underline{V}}^{(i, m+\frac{1}{2})} - \left(\underline{\underline{I}}_3 \otimes \underline{\underline{C}} \right) \underline{\underline{T}}_T^{(\Gamma ik, m+\frac{1}{2})} \quad (30)$$

where

$$\underline{\underline{C}} = \frac{\delta^\Gamma \Delta t S^{ik} (\rho^i W^i + \rho^k W^k)}{\rho^i \rho^k W^i W^k} \underline{\underline{K}}^{e-1} \underline{\underline{F}}^e, \quad (31)$$

and

$$\underline{\underline{V}}^{(i, m+\frac{1}{2})} = \underline{\underline{V}}^{(i, m-\frac{1}{2})} + \underline{\underline{R}}_T^{(i, m)} - \underline{\underline{R}}_T^{(k, m)} + \left(\underline{\underline{I}}_3 \otimes \underline{\underline{C}} \right) \underline{\underline{T}}_T^{(\Gamma ik, 0)}, \quad (32)$$

is the slip rate assuming a frictionless fault free surface, so it will be called *free slip rate*. Since the total tangential traction magnitude must equal the fault's shear strength, eq.(10), it should be computed as

$$\left\| \underline{\underline{T}}_T^{(\Gamma ik, m+\frac{1}{2})} \right\| = \tau^{(\Gamma ik, m+\frac{1}{2})} = f \left(\underline{\underline{V}}^{(i, m+\frac{1}{2})}, \underline{\underline{\psi}}^{(i, m+\frac{1}{2})} \right) \sigma_N^{(i, m)}. \quad (33)$$

Following a staggered time stepping of the velocity-state system, as done by Rojas *et al.* (2009), we decouple the state variable from the previous equation computing the friction coefficient with the state variable and the stresses simultaneously

$$\tau^{(\Gamma ik, m+\frac{1}{2})} = f \left(\underline{\underline{V}}^{(i, m+\frac{1}{2})}, \underline{\underline{\psi}}^{(i, m)} \right) \sigma_N^{(i, m)}. \quad (34)$$

So the state variable can be updated independently for each degree of freedom (DOF) located in the fault following the next steps:

1. Update the low-velocity steady state friction coefficient, eq. (16),

$$f_{LV}^{(i, m)} \left(\underline{\underline{V}}^{(i, m-\frac{1}{2})} \right) = f_0^i - (b^i - a^i) \ln \left(\underline{\underline{V}}^{(i, m-\frac{1}{2})} / V_0^i \right). \quad (35)$$

2. Compute the steady state friction coefficient, eq. (15),

$$f_{ss}^{(i,m)} \left(V^{(i,m-\frac{1}{2})} \right) = f_w^i + \frac{f_{LV}^{(i,m)} \left(V^{(i,m-\frac{1}{2})} \right) - f_w^i}{\left[1 + \left(V^{(i,m-\frac{1}{2})} / V_w^i \right)^8 \right]^{1/8}}. \quad (36)$$

3. Update the steady state variable, eq. (14),

$$\psi_{ss}^{(i,m)} \left(V^{(i,m-\frac{1}{2})} \right) = a^i \cdot \ln \left\{ \frac{2V_0^i}{V^{(i,m-\frac{1}{2})}} \sinh \left[\frac{f_{ss}^{(i,m)} \left(V^{(i,m-\frac{1}{2})} \right)}{a^i} \right] \right\}. \quad (37)$$

4. Update the state variable with the exact solution of its ODE, eq. (13),

$$\begin{aligned} \psi^{(i,m)} \left(V^{(i,m-\frac{1}{2})} \right) = & \psi^{(i,m-1)} \left(V^{(i,m-\frac{3}{2})} \right) \exp \left(-\frac{V^{(i,m-\frac{1}{2})} \Delta t}{L} \right) \\ & + \psi_{ss}^{(i,m)} \left(V^{(i,m-\frac{1}{2})} \right) \left(1 - \exp \left(-\frac{V^{(i,m-\frac{1}{2})} \Delta t}{L} \right) \right). \end{aligned} \quad (38)$$

In order to compute the total tangential traction, $\underline{T}_T^{(\Gamma ik, m+\frac{1}{2})}$, the slip-rate norm, $V^{(i, m+\frac{1}{2})}$ must be computed first. To do so, we must first recognize that the slip, as well as the free slip, should be parallel to the total tangential traction due to the slip direction condition, eq. (17), so we can write the slip rate update scheme, eq. (30), as

$$\underline{V}^{(i, m+\frac{1}{2})} = \left(\underline{I}_3 \otimes \underline{L} \right) \underline{V}^{(i, m+\frac{1}{2})} - \left(\underline{I}_3 \otimes \underline{CN} \right) \underline{V}^{(i, m+\frac{1}{2})}, \quad (39)$$

where $\underline{L} = \text{Diag} \left(\underline{\hat{V}}_M^{(i, m+\frac{1}{2})} \odot \underline{V}_M^{-1(i, m+\frac{1}{2})} \right)$, $\underline{N} = \text{Diag} \left(\underline{T}_{TM}^{(\Gamma ik, m+\frac{1}{2})} \odot \underline{V}_M^{-1(i, m+\frac{1}{2})} \right)$, $\underline{\hat{V}}_M^{(i, m+\frac{1}{2})}$ is a vector with the norm of the free slip rate of all DOFs of the i -tetrahedron, $\underline{V}_M^{-1(i, m+\frac{1}{2})}$ is a vector with the inverse of the norm of the slip rates of all DOFs of the i -tetrahedron, $\underline{T}_{TM}^{(\Gamma ik, m+\frac{1}{2})}$ is a vector with the norm of the total tangential tractions of all DOFs of the i -tetrahedron and $\odot$ is the operator for the component-wise multiplication.

Substituting the total tangential traction with the Amon-ton-Coulomb model, eq. (10), and the friction coefficient, eq. (12), in eq. (39), it can be reformulated as a non-linear equation, $\underline{r}(\cdot)$, that depends only on the magnitudes of the slip rates, $\underline{V}_M^{(i, m+\frac{1}{2})}$, as

$$\underline{r} \left(\underline{V}_M^{(i, m+\frac{1}{2})} \right) := \underline{V}_M^{(i, m+\frac{1}{2})} + \underline{C} \left(\underline{a}^i \cdot \text{asinh} \left[\frac{V_M^{(i, m+\frac{1}{2})}}{2V_0^i} \exp \left(\frac{\psi^{(i, m)}}{\underline{a}^i} \right) \right] \underline{a}_N^{(i, m)} \right) - \underline{\hat{V}}_M^{(i, m+\frac{1}{2})} = 0, \quad (40)$$

which is solved with a Newton-Raphson method to update the slip rate in every p -iteration

$$\underline{V}_M^{(i, m+\frac{1}{2})^{p+1}} = \underline{V}_M^{(i, m+\frac{1}{2})^p} - \underline{J}^{-1} \underline{r} \left(\underline{V}_M^{(i, m+\frac{1}{2})^p} \right), \quad (41)$$

where $\underline{J}$ is the Jacobian, which is defined in [Appendix C](#).

Once the Newton-Raphson method has converged, the total tangential traction is computed as

$$\underline{T}_T^{(\Gamma ik, m+\frac{1}{2})} = \underline{f} \left(\underline{V}_M^{(i, m+\frac{1}{2})}, \underline{\psi}^{(i, m)} \right) \odot \underline{a}_N^{(i, m)} \odot \underline{\hat{V}}^{(i, m+\frac{1}{2})}, \quad (42)$$

where $\underline{\hat{V}}^{(i, m+\frac{1}{2})}$ is the normalized free slip.

Finally, to update the velocity in eq. (28), we evaluate the tangential traction, $\underline{T}_T^{(\Gamma ik, m+\frac{1}{2})}$ using the total tangential traction, $\underline{T}_{TM}^{(\Gamma ik, m+\frac{1}{2})}$, with eq. (42).

3.2. Fault normal traction, $\underline{T}_N^{(\Gamma ik, m+\frac{1}{2})}$, computation

For the computation of the fault normal traction, $\underline{T}_N^{(\Gamma ik, m+\frac{1}{2})}$, we force the continuity of the normal component of the velocity to avoid an opening mode during rupture, as

$$\underline{V}_N^{(i, m+\frac{1}{2})} = \underline{v}_N^{(i, m+\frac{1}{2})} - \underline{v}_N^{(k, m+\frac{1}{2})} = 0. \quad (43)$$

We project the velocity update scheme, eq. (26), along the fault normal and substitute it in eq. (43) to clear the normal fault traction as

$$\underline{T}_N^{(\Gamma ik, m+\frac{1}{2})} = \frac{\delta \Gamma \rho^i \rho^k W^i W^k}{\Delta t S^{ik} (\rho^i W^i + \rho^k W^k)} \left(\underline{I}_3 \times \underline{K}^e \underline{F}^{e-1} \right) \left[-\underline{R}_N^{(i, m)} + \underline{R}_N^{(k, m)} \right]. \quad (44)$$

3.3. Artificial viscosity to attenuate numerical noise

The abrupt stress drop at the rupture front induces numerical noise in a high frequency band (HFO). This issue has been observed in diverse numerical methods, such as FDM (Day, 1982; Cruz-Atienza and Virieux, 2004; Day *et al.* 2005), FEM (Oglesby *et al.*, 1998; Barall, 2009; Aagaard *et al.*, 2013), FVM (Benjemaa *et al.*, 2009) and DGM (Tago *et al.*, 2012). Some DGM strategies have used upwind fluxes to solve the Riemann problem due to the velocity discontinuity along the fault, and because of their dissipative nature, they do not generate HFOs (De la Puente *et al.*, 2009; Pelties *et al.*, 2012; Zhang *et al.*, 2023). However, the amount of dissipation and dispersion introduced depends strongly on the discretization and order of approximation. Instead of using an upwind flux, whose dissipation

pative effect cannot be controlled, we opted for a simple, but very efficient, strategy used by Day (1982), and in subsequent papers, and described by Day *et al.* (2005), that introduces an artificial viscosity with a damped stress term for the motion equation, (1), as

$$\rho \partial_t \underline{v} = \sum_{\theta \in \{x,y,z\}} \partial_{\theta} \underline{M}_{\theta} (\underline{\sigma} + \gamma \partial_t \underline{\sigma}), \quad (45)$$

where γ is an artificial damping parameter.

Rojas *et al.* (2009) made a detailed analysis of this damping parameter, γ , and showed that for a given discretization and small values of γ , the absorption-wavelength spectrum can be far separated from the cohesive zone dimension. This means that the physical wavelength spectrum is practically unaffected and only the HFOs are eliminated. This strategy is appealing for energy conservative numerical schemes like ours (due to the central fluxes and the ad hoc fault fluxes). The damping parameter, γ , will be discussed in the following section.

The DGM implementation of the artificial damping is straightforward; instead of using $\underline{\sigma}^{(i,m)}$ in the velocity update scheme, eq. (21), we use the damped stresses, $\underline{\sigma}_{\text{damped}}^{(i,m)} = \underline{\sigma}^{(i,m)} + \gamma \frac{\underline{\sigma}^{(i,m)} - \underline{\sigma}^{(i,m-1)}}{\Delta t}$.

4. Results

The verification and performance assessment of our numerical implementation was done by solving the benchmark TPV104 from SCEC (Harris *et al.*, 2018).

4.1. TPV104 SCEC benchmark

The TPV104 is a strike-slip planar fault embedded in an isotropic, homogeneous linearly elastic half-space that reaches the free surface. The hypocenter is located in the middle of a velocity-weakening portion of dimensions of 30 km $\times$ 15 km. At its borders, there is a 3 km zone of transition from velocity-weakening to velocity-strengthening, beyond which the fault is velocity-strengthening. The values of all the friction-law parameters, initial conditions and the nucleation parameters can be found in the benchmark website https://strike.scec.org/cvws/tpv103_104docs.html. The locations of the nine on-fault stations and six free-surface stations can also be consulted.

We use Gmsh software (Geuzaine and Remacle, 2009) to discretize the domain with a tetrahedral conforming mesh. The fault plane is defined before the mesh generation, such that it cannot be crossed by any tetrahedron. For subsections 4.2 and 4.3, we discretized the fault with an edge-length characteristic

size of 100 *m*. Since we use second-order elements, the distance between DOFs along the edge length is 50 *m*.

We start by evaluating the effect of the artificial damping parameter, γ , on dispersion and accuracy of our solution. Then, due to the absence of an analytical solution, we compare our results with those obtained using other numerical strategies. Finally, we evaluate the convergence as a function of grid refinement.

4.2. Artificial damping

As clearly stated by Rojas *et al.* (2009), the larger the value of γ , the smaller the amplitude of the HFOs. However, large γ can also lead to overdamped solutions and a significant delay in the rupture front.

Figure 1 shows the shear stress and slip rate using the damping factors $\gamma = \{0.0, 0.1, 0.2, 0.3, 0.4\}$, registered on one of the farthest on-fault stations located 12 km along strike and 12 km along dip from the hypocenter. From Figures 1.a and 1.d, we do not observe a significant difference for those γ values in the shear stress and slip rate, respectively. To better appreciate the difference in the shear stress, we made zooms around the peak shear stress, Figure 1.b, and the noisiest time interval, Figure 1.c, respectively. The maximum per cent difference of the peak shear stress with respect the non-damping solution, $\gamma = 0.0$, was of 0.0709% for $\gamma = 0.4$. And also for $\gamma = 0.4$, the maximum percent difference in the delay between peak shear stresses, relative to the undamped solution, is 0.2672%. When assessing the noisiest time interval, we find a strong reduction of the HFOs. Since for $\gamma = 0.1$ we still observe small high frequency oscillations, we choose $\gamma = 0.2$ as an optimal value. For the slip rate, the HFOs were eliminated for all $\gamma > 0$ (see Figure 1.f). As illustrated in Figure 1.e, the peak slip rate differences are almost negligible; the maximum percent difference relative to the undamped solution was 1.1141% for $\gamma = 0.4$ and for $\gamma = 0.2$ it was 0.4093%. It is important to state that optimal γ is *h*-dependent, however Rojas *et al.* (2009) showed that for $\gamma \in (0, 0.1]$ the HFOs are satisfactorily removed without perturbing the important features of the undamped solution.

Figure 2 shows the convergence results with common selected rupture features to assess the accuracy (Day *et al.*, 2005; Pelties *et al.*, 2012; Zhang *et al.*, 2023). The damped solutions for $\gamma = \{0.1, 0.2, 0.3, 0.4\}$ are compared with the undamped solution, $\gamma = 0.0$, by computing the Root Mean Squared Percentage Error (RMSPE) of the rupture time along the fault (Figure 2.a), the RMSPE peak slip rate (Figure 2.b) and the RMSPE final slip (Figure 2.c), with the latter two considering all the on-fault stations of the benchmark. The convergence rates are 0.4952 for the rupture time, 1.0382 for the peak slip rate, and 0.7324 for the final slip. All convergence rates in the range of $\gamma = [0.1, 0.4]$

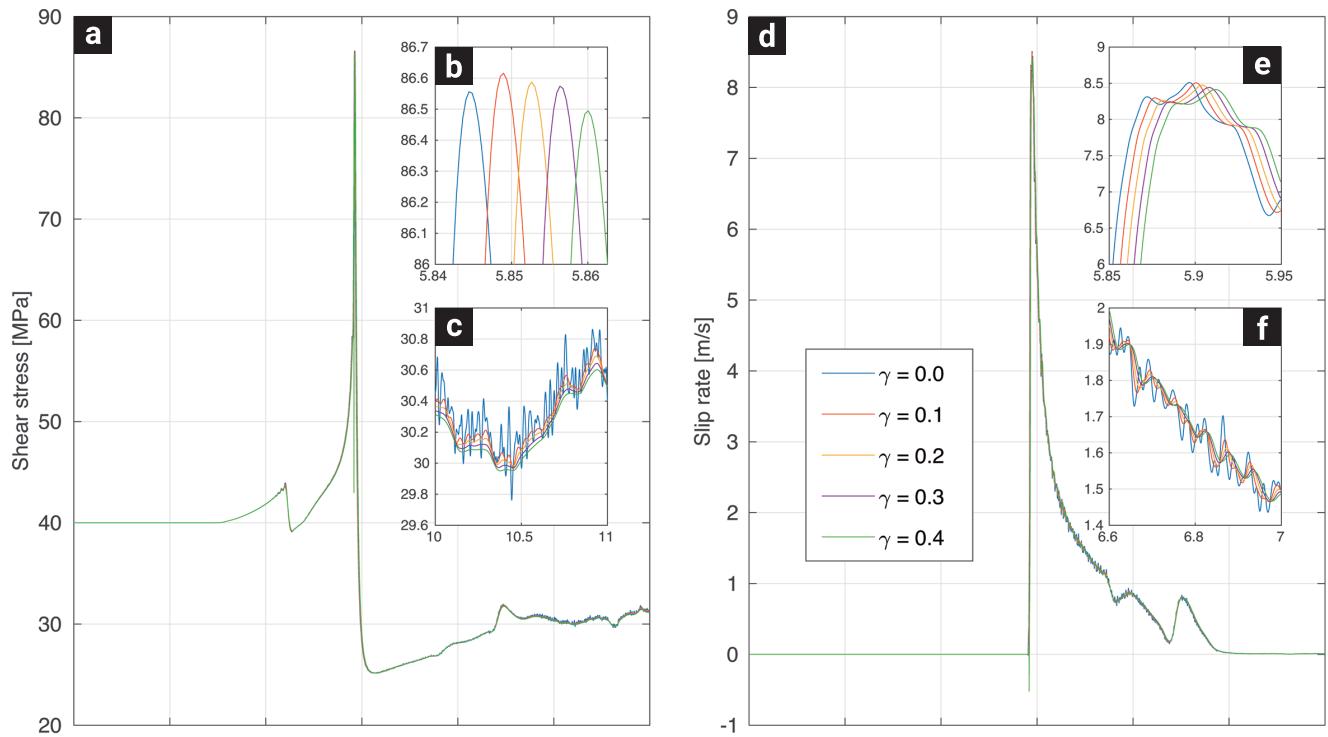


Figure 1. On-fault records for damping factors, $\gamma = \{0.0, 0.1, 0.2, 0.3, 0.4\}$, of the shear stress [MPa], (a), and slip rate [m/s], (d), at the fault station located 12 km along strike and 12 km along dip from the hypocenter. (b) and (c) are zoom figures of the peak shear stress and the noisiest shear stress time interval, respectively. (e) and (f) are zoom figures of the peak slip rate and the noisiest slip rate time interval, respectively.

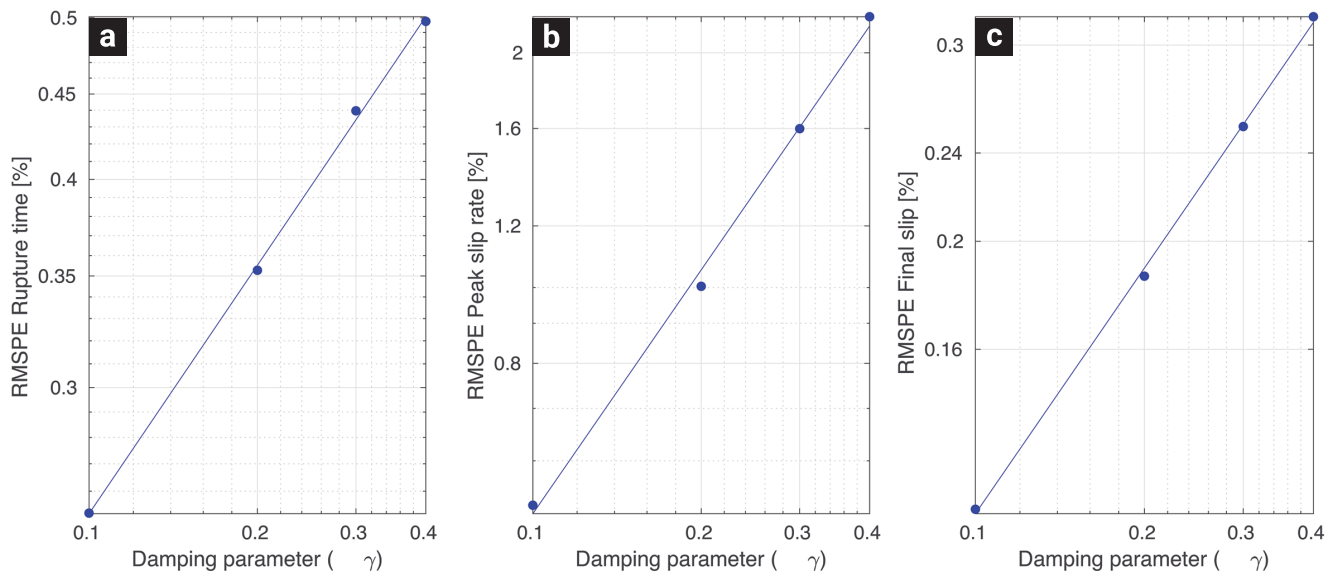


Figure 2. Convergence results of TPV104 for $\gamma = 0.1, 0.2, 0.3, 0.4$ relative to the un-damped solution, $\gamma = 0.0$. The circles are the RMSPE for selected rupture features and the lines are the logarithm linear regressions. (a) RMSPE of the rupture time along the fault with convergence rate of 0.4952. (b) RMSPE of the peak slip rate considering all the on-fault stations of the benchmark with convergence rate of 1.0382. (c) RMSPE of the final slip considering all the on-fault stations of the benchmark with a convergence rate of 0.7324.

are small as expected from the previous analysis of Figure 1. So even with large values of γ , compared to the recommendation of $\gamma < 0.1$, we still preserve the main features of the rupture propagation. The peak slip rate is the most sensitive to the damping parameter, so it would be an adequate feature to compare with other numerical solutions. Our solutions in the next subsection were computed with our selected optimal damping factor $\gamma = 0.2$, whose RMSPE values for rupture time, peak slip rate, and final slip are 0.3528, 1.0040, and 0.1861, respectively.

4.3. Waveform assessment

Since no analytical solutions exist for dynamic rupture problems, we selected three numerical approaches for comparison with our solution. These reference solutions were computed with a roughly comparable DOF density in order to illustrate the variability among similar methods. One exception was deliberately oversampled by a factor of four to serve as a high-precision benchmark. It is important to note that equalizing the DOF density should not be interpreted as a basis for ranking the methods, as DOF density is not directly related to computational cost.

The numerical schemes selected for comparison were FaultMod, a FEM code (Barall, 2009); SPECFEM3D, a SEM code (Ampuero, 2002; Kaneko *et al.*, 2008); and MixF-DG, a DGM code (Zhang *et al.*, 2023). FaultMod is a first-order FEM code that employs Newmark time integration, artificial damping to mitigate HFOs and a grid-doubling technique for mesh refinement near the fault. An element size of 50 *m* was used for TPV104, corresponding to the recommended spacing between DOFs for this benchmark. This strategy has been used as a reference for all SCEC benchmarks since it was developed by the same group in charge of SCEC benchmarks development. SPECFEM3D is SEM with an adapted Newmark time integration and without artificial damping. The fault element size was also 50 *m* with 5 Gauss-Lobato-Legendre (GLL) nodes per edge for fourth order elements. The distribution of the GLL nodes is not equidistant, but we consider a DOF's equivalent distance of 12.5 *m*. Finally, MixF-DG is the most recent DGM reported to date with upwind fluxes everywhere except on the off-fault fluxes of the on-fault elements where central fluxes are implemented. It does not add artificial damping due to the upwind on fault flux that does not generate HFOs. The use of central fluxes prevents spatial spike oscillations (SSO) generated by large on-fault stress drops and the use of upwind fluxes. These SSOs are strongly mesh-dependent and can lead to instabilities, specially with near-fault mesh refinement. The fault element size was 200 *m* with fourth order elements as well. The node distribution is optimal and not equidistant (Hesthaven and Warburton, 2007), but we consider the DOFs equivalent distance of 50 *m*. For DGCrack we use second

order elements with a fault element size of 100 *m*, i.e. a 50 *m* DOFs equivalent distance. Since, of all strategies, SPECFEM3D has the smallest DOFs distance and it does not add any damping procedure, explicitly nor implicitly, it was selected as our reference solution. FaultMod and SPECFEM3D discretize the medium using hexahedral elements, whereas MixF-DG and DGCrack use tetrahedral elements.

In Figure 3, the rupture velocity along the fault is in good agreement among all solutions except for FaultMod solution, where a rupture delay can clearly be observed at the edges. On-fault records of the shear stress and the slip rate are shown in Figure 4. On-fault records located 9 km along strike and 7.5 km along dip from the hypocenter are shown in Figures 4.a and 4.b. Those located 12 km along strike and 3 km along dip from the hypocenter are shown in Figures 4.e and 4.g. All of them are quite similar, with a small delay of the FaultMod solution, as was expected from the rupture time contour plots (Figure 3). This delay can be better appreciated in the zoom insets around the peaks of the shear stress and the slip rate, Figures 4.f and 4.h, respectively. The SPECFEM3D solution is the only one that we expect to have no delay since it has the smallest DOF spacing and no damping was applied. Afterwards, the order of the rupture front is DGCrack, MixF-DG and FaultMod, which can be appreciated in the peak shear stress delay in Figure 4.f. Thus, with our (explicitly imposed) artificial damping, the dispersion is smaller than that produced by the implicit damping of the upwind flux from MixF-DG. However, artificial damping requires the specification of a hyperparameter, γ , which must be tuned through numerical tests or heuristic rules. Comparing the peak slip rate at a single station is challenging, as the differences may be more related to interpolation effects than to attenuation artifacts. Figure 4.h shows that MixF-DG has the lowest amplitude, but it would be better to assess the differences considering all fault stations (Figure 6.b).

The noise in the on-fault records seems negligible for all numerical approaches (see Figure 4.c). However, this is not the case for the off-fault records. Figure 5 shows the velocity field registered at a free surface station located 12 km along strike and 6 km perpendicular to the fault. It should be noticed that not all noise comes from the strong singularity evolution of the rupture front, but it can also be due to the mesh coarsening used by the different strategies for better computational performance. Unfortunately, the noise differentiation for the different strategies is not possible, so we will assume that most of the noise comes from the dynamic rupture modeling. All numerical approaches show noise in the records but with different frequency content. As it is expected, SPECFEM3D has the largest peak velocities (Figures 5.b, 5.d and 5.f) but also the largest HFOs. In amplitude, our solution, DGCrack, seems to be closest to SPECFEM3D.

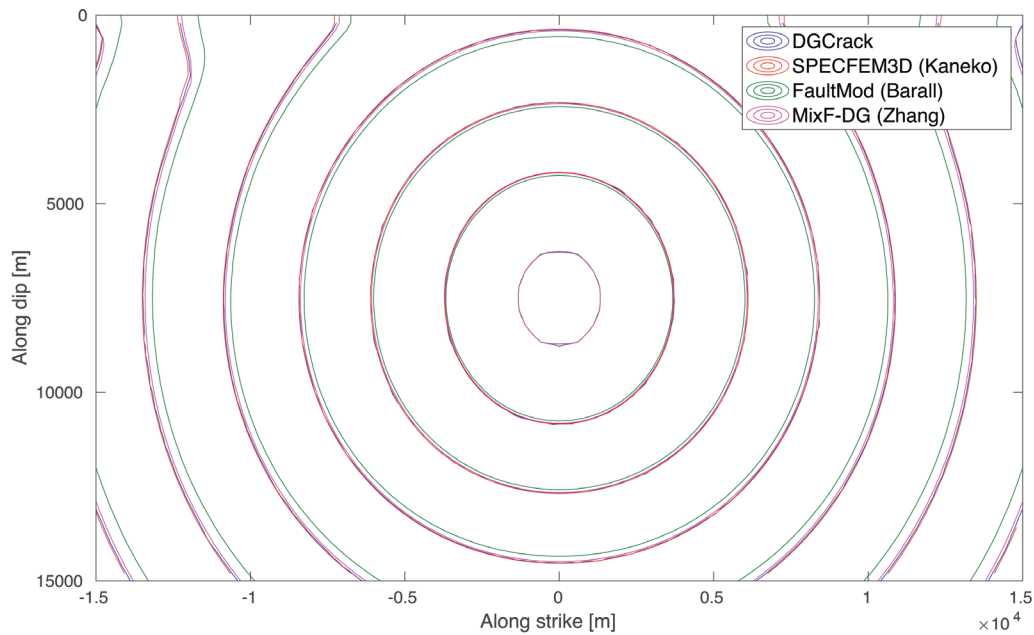


Figure 3. TPV104 Contour plots of rupture times along the fault.

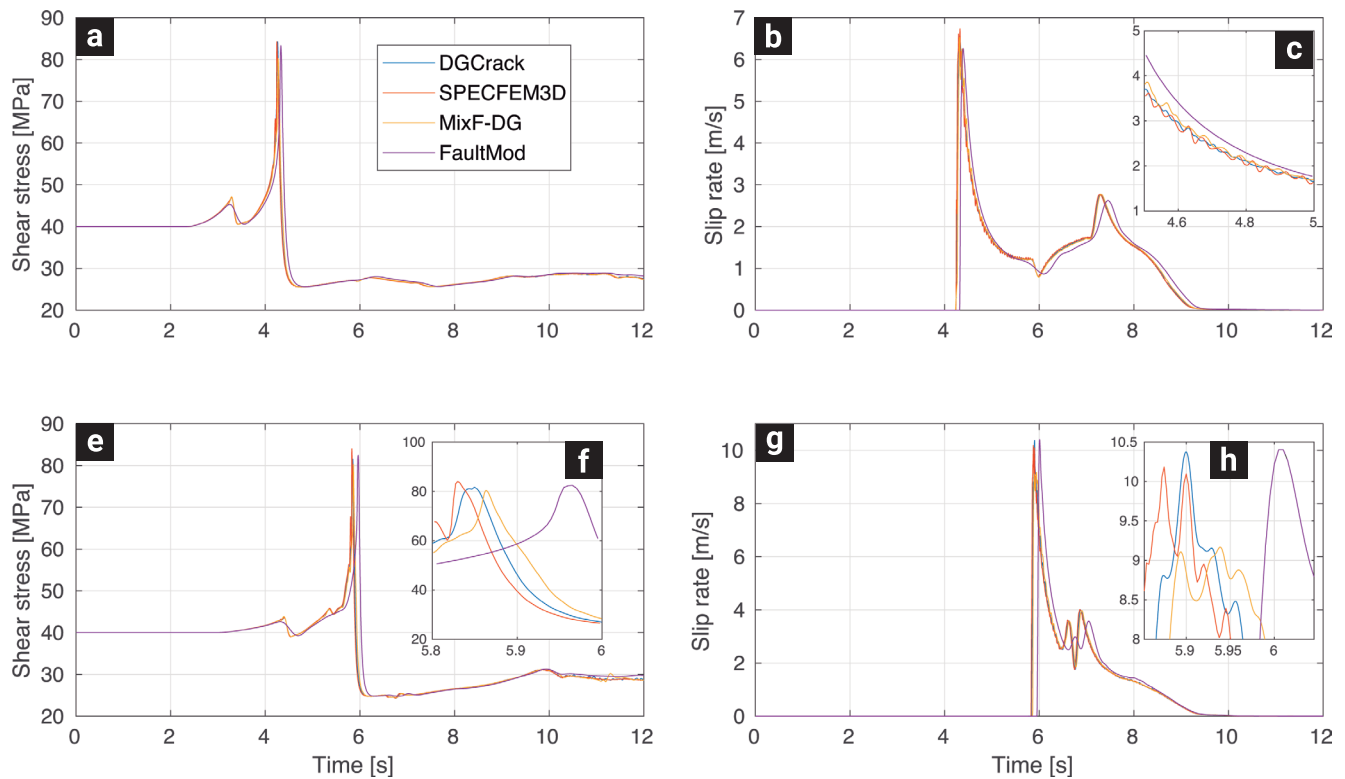


Figure 4. On-fault records for the different numerical strategies, DGCrack, SPECFEM3D, MixF-DG and FaultMod. The DOFs equivalent distance for DGCrack, MixF-DG and FaultMod is 50 m, while for SPECFEM3D is 12.5 m. Horizontal shear stress [MPa] and slip rate [m/s] at the fault station located 9 km along strike and 7.5 km along dip from the hypocenter, (a) and (b) respectively. (c) shows the noisiest segment of the slip rate plot in (b). Horizontal shear stress [MPa] and slip rate [m/s] at the fault station located 12 km along strike and 3 km along dip from the hypocenter, (e) and (g) respectively. (f) and (h) are plots around the peak of the shear stress, (e), and the slip rate, (g), respectively.

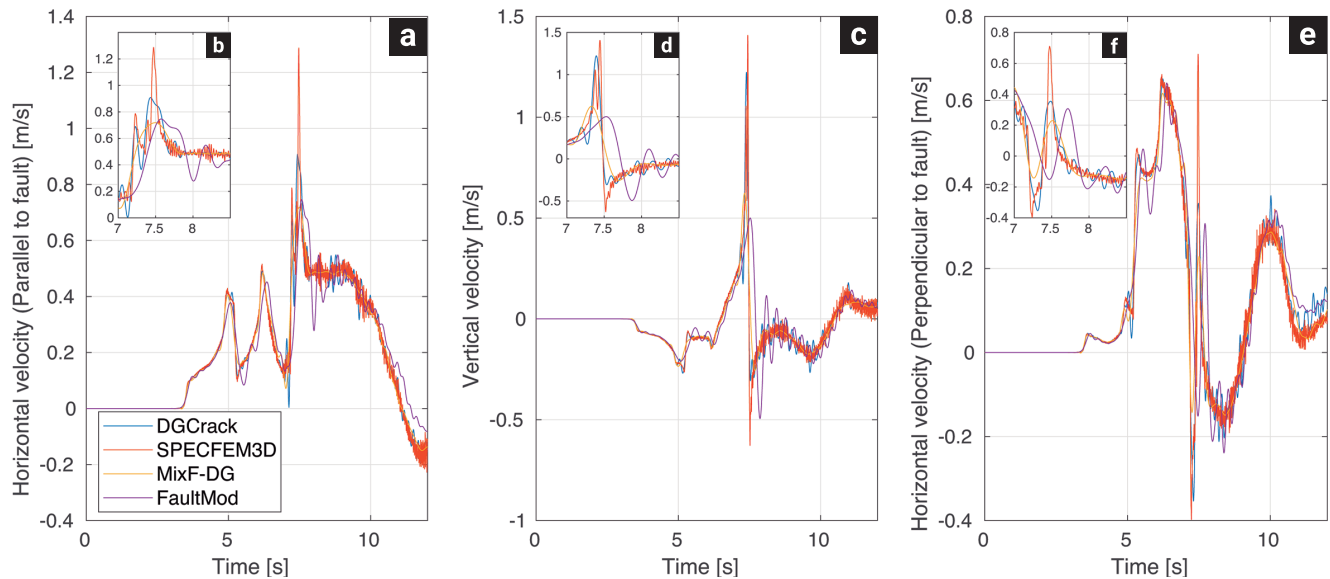


Figure 5. Free surface off-fault records at a station located 12 km along strike and 6 km perpendicular to the fault. The DOFs equivalent distance for DGCrack, MixF-DG and FaultMod is 50 m, while for SPECFEM3D is 12.5 m. (a) and (e) are the parallel and perpendicular to the fault horizontal velocity field components [m/s] and (c) is the vertical velocity field component [m/s]. (b), (d) and (f) show the intervals with the peak values of the corresponding velocity field.

FaultMod and MixF-DG seems to have similar peak velocities with even smaller values. It should be noticed that MixF-DG shows the cleanest records and the same phase arrivals as the rest of the solutions.

Figure 6 shows the RMSPE for the on-fault metrics used by Zhang *et al.* (2023), which includes the rupture time along the fault and the peak slip rate and the final slip recorded at nine on-fault stations. The RMSPEs are computed using the SPECFEM3D solution as the reference, due to its finer fault discretization. We can observe that for all these metrics DGCrack is closest to SPECFEM3D, followed closely by MixF-DG. In every metric, DGCrack achieves a RMSPE roughly 50% lower than MixF-DG's. FaultMod shows the worst RMSPEs, but as shown in Figure 4, the on-fault records are almost noise-free and similar to the other solutions.

We have shown that DGCrack's solution is in agreement with other numerical strategies. In the next section we will show its convergence under grid refinement.

4.4. Convergence under grid refinement

In this section, we assess the accuracy according to the grid size. We solved TPV104 for different element sizes, $h = \{100, 150, 200, 300, 400\}$ m, with the previously selected artificial damping parameter $\gamma = 0.2$.

Figure 7 shows that the horizontal shear stress and the slip rate for the different discretizations are quite similar even

with the discretization of $h = 300$ m that is three times larger than the suggested discretization for the cohesive zone. Figure 7.c shows low frequency noise that seems to get stronger with grid coarsening. A slight delay in the rupture front is more clearly visible in the zoomed views around the peak shear stress and slip rate (Figures 7.f and 7.h, respectively). However, it is remarkable how the peak slip rate remains nearly identical across all discretizations.

The convergence rates are shown in Figure 8 as a function of grid refinement. The convergence rates for the rupture time, the peak slip rate and the final slip, were 1.6008, 0.6774 and 1.4470, respectively. These low convergence rates are expected due to the nonlinearity of the friction equations. Moreover, higher-order approximations do not improve the convergence rates, as has been shown numerically (Zhang *et al.*, 2023; Pelties *et al.*, 2012).

5. Discussion and Conclusions

We have implemented a rate-and-state friction law in DGCrack (Tago *et al.*, 2012), inspired by the work of Rojas *et al.* (2009). The friction coefficient computation is done with a staggered time strategy for the velocity-state system. Unlike previous implementations with DGM (e.g. Pelties *et al.*, 2012; Zhang *et al.*, 2023), we do not employ an upwind flux; instead, we use an ad hoc flux along the fault. While an upwind flux

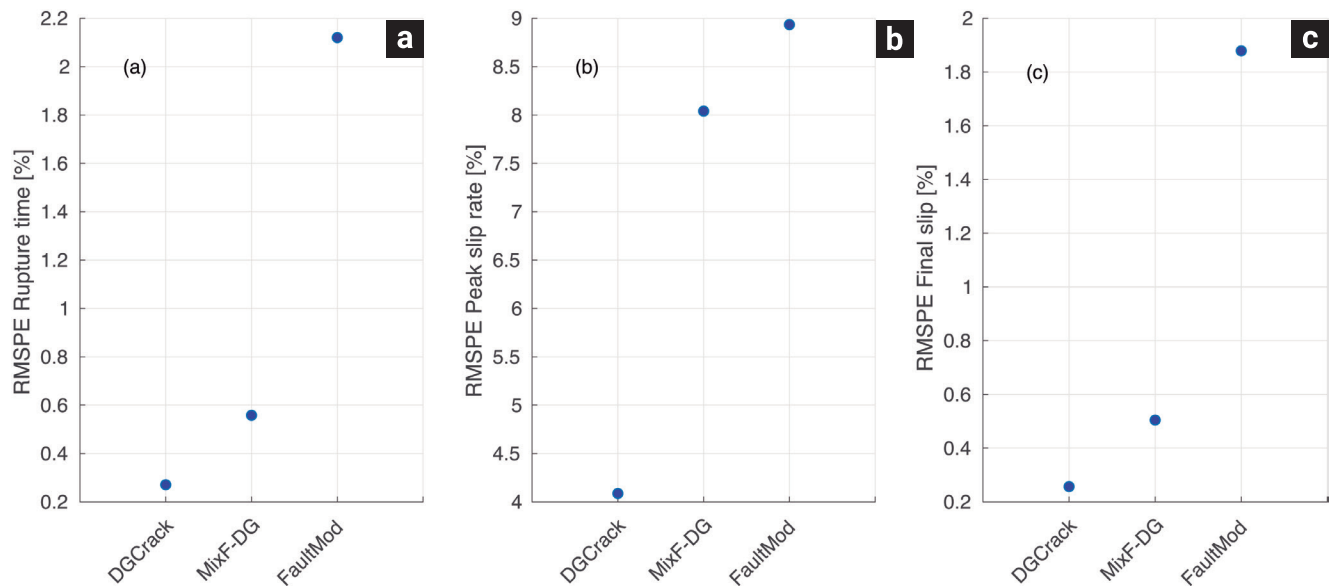


Figure 6. On-fault metrics comparison with SPEC3D. (a) RMSPE Rupture time [%] along the fault. (b) RMSPE Peak Slip rate [%] considering all 9 on-fault records. (c) RMSPE Final slip [%] considering all 9 on-fault records.

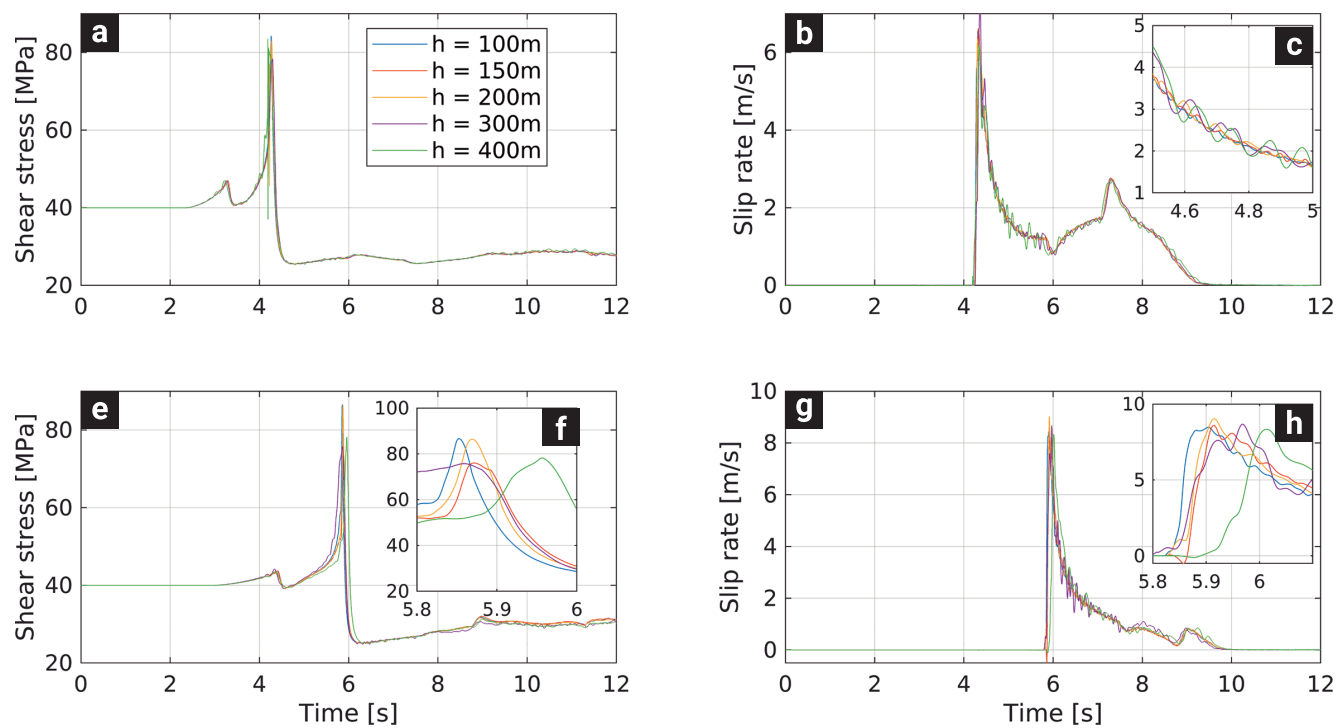


Figure 7. On-fault records for different grid sizes, $h = \{100, 150, 200, 300, 400\}$ m. Horizontal shear stress [MPa] and slip rate [m/s] at the fault station located 9 km along strike and 7.5 km along dip from the hypocenter, (a) and (b) respectively. (c) shows the noisiest segment of the slip rate plot in (b). Horizontal shear stress [MPa] and slip rate [m/s] at the fault station located 12 km along strike and 12 km along dip from the hypocenter, (e) and (g) respectively. (f) and (h) are plots around the peak of the shear stress, (e) and the slip rate, (g), respectively.

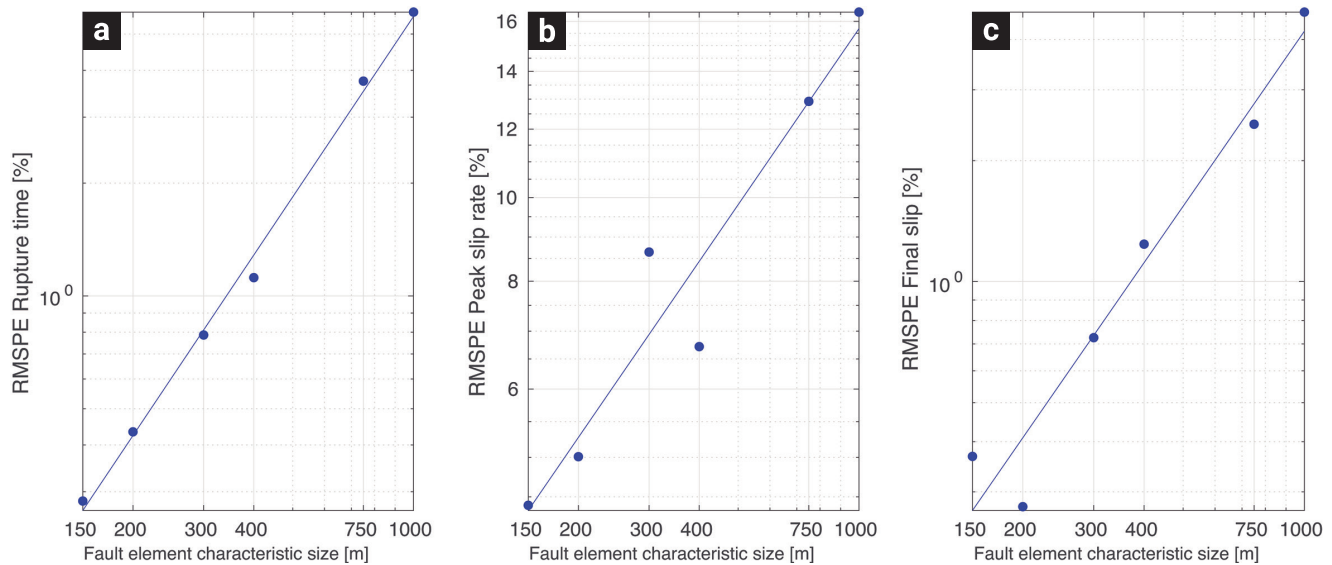


Figure 8. Convergence results of TPV104 for $h = \{150, 200, 300, 400, 750, 1000\}$ m relative to the finest grid size, $h = 100$ m with $\gamma = 0.2$. The circles are the RMSPE for selected rupture features and the lines are the logarithm linear regressions. (a) RMSPE of the rupture time along the fault with convergence rate of 1.6008. (b) RMSPE of the peak slip rate considering all the on-fault stations of the benchmark with convergence rate of 0.6774. (c) RMSPE of the final slip considering all the on-fault stations of the benchmark with a convergence rate of 1.4470.

naturally suppresses HFOs due to its intrinsic dissipation, the magnitude of the dissipation and dispersion introduced has not been studied in detail.

Our ad hoc fault flux generates HFOs, which we attenuated using an artificial damping strategy. This required the introduction of a damping parameter that needs tuning. As previously shown by Rojas *et al.* (2009), for $\gamma < 0.1$ the damped frequency band lies above the frequency band in which the undamped numerical scheme has effectively converged, so almost no unwanted damping is applied. In practice, the phase delay produced by the artificial damping is more noticeable than any amplitude reduction. However, as illustrated in Figures 4 and 2, these effects are negligible. For the TPV104 case, we chose a larger damping parameter than the recommended value, $\gamma = 0.2$, to eliminate all visible noise. For earthquake simulations, however, we recommend starting with a conservative value of $\gamma = 0.1$, increasing it only if HFOs persist.

In the waveform comparison, DGCrack demonstrated good performance relative to recent and validated numerical strategies. For the fault metrics shown in Figure 6, our solution was closer to the reference than that of MixF-DG, even though MixF-DG employed a higher-order approximation. Furthermore, dissipation and dispersion in DGCrack fault fields were smaller than in MixF-DG fault fields (see Figures 4.f and 4.h). These numerical artifacts appear to arise more from the upwind flux used in MixF-DG than from the artificial damping introduced in our approach.

Convergence rates from grid refinement are inherently limited by the nonlinearity of dynamic rupture modeling. For DGCrack, the convergence rates obtained are relatively low, as expected from the earlier numerical comparison of the SCEC slip-weakening TPV3 benchmark by Zhang *et al.* (2023). This should not be viewed as a drawback, since our solution delivers accuracy comparable to higher-order approximations. Moreover, for coarser grids DGCrack should outperform high-order numerical strategies due to its relatively low convergence rates (see Figure 7.h).

Although we have shown that for $\gamma < 0.1$ the fault field spectra are largely preserved, the need to introduce a damping parameter may be seen as a disadvantage compared with approaches that use upwind fluxes to avoid HFOs. However, since numerical damping is necessary, it may be better to have a control parameter for it, rather than it being implicit in a numerical flux. We believe that an upwind flux formulation with controlled dissipation and dispersion still needs to be developed to address this limitation. Finally, we plan to apply DGCrack to simulate earthquakes in Mexico, using friction parameters and initial conditions inferred from seismotectonic studies (Cruz-Atienza *et al.*, 2021).

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Data Availability Statement: The numerical solutions used for the comparison can be downloaded at <https://strike.scec.org/cvws/cgi-bin/cvws.cgi>. The DGCrack solution is available from the authors upon request.

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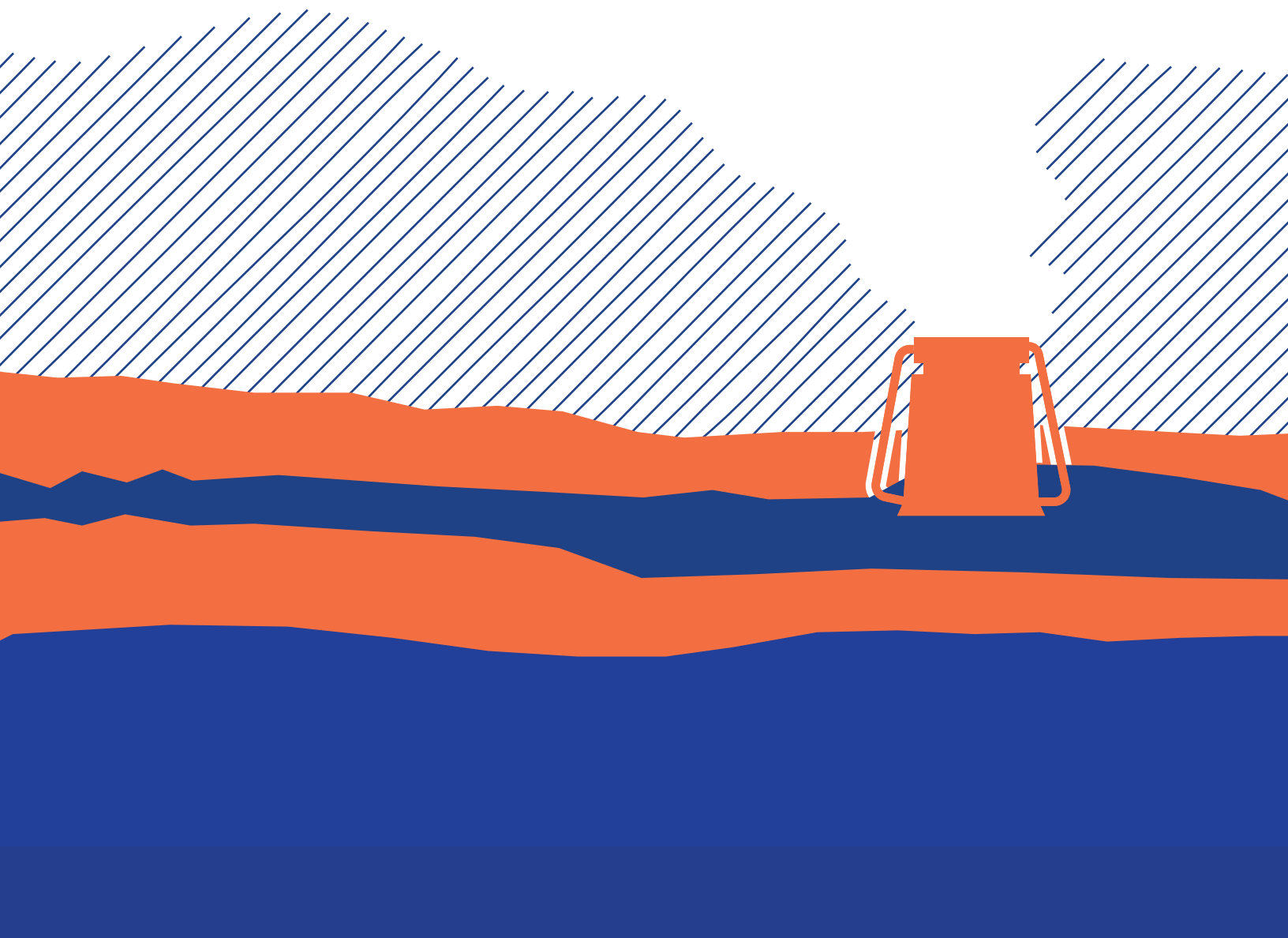
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Sección especial

Vulcanismo y Geotermia

Special section

Volcanism and Geothermics



Geophysical model of the hydrothermal system of the Amado Nervo Volcano (Nayarit, Mexico) and preliminary evaluation of its geothermal potential

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Abstract

This study presents a conceptual model of the hydrothermal system associated with the Amado Nervo Volcano in western Mexico, integrating geological, gravity, magnetic, and geothermal data. The objectives were to characterize the subsurface structure of the volcanic system, identify geophysical signatures potentially associated with hydrothermal alteration, and provide a preliminary estimate of the region's exploitable geothermal potential. Aeromagnetic data were processed using reduction to the pole, analytical signal, and upward continuation, and spectral analysis techniques, whereas gravity data were derived from the Global Gravity Model (GGM+) and transformed into simple and residual Bouguer anomalies. The radially averaged power spectrum method was applied to estimate magnetic source depths, the Curie point depth, the geothermal gradient, and conductive heat flow. The geophysical results reveal a special association among magnetic attenuation zones, analytical signal minima, and negative gravity anomalies, delineating structurally complex domains characterized by reduced density and magnetization. Spectral analysis yielded a Curie point depth of approximately 7.5 km, from which a geothermal gradient of 70.8 °C·km⁻¹ and a conductive heat flow of 176 mW·m⁻² were estimated. These values are consistent with those reported for geothermal regions of the Trans–Mexican Volcanic Belt. Integration of geophysical and geological information enabled the development of a conceptual hydrothermal model of a structurally controlled volcanic–hydrothermal system potentially sustained by a residual intrusive heat source. A preliminary geothermal resource assessment was conducted using the volumetric Heat in Place method coupled with 9,999 Monte Carlo simulations. The results yielded a mean geothermal potential of 19.75 MW, with simulated values ranging from approximately 5 to 34 MW. Although subject to uncertainties in reservoir geometry and thermophysical parameters, the estimated resource ranks the Amado Nervo volcanic system among medium–scale geothermal prospects in Mexico. The proposed hydrothermal architecture shows similarities to that of the nearby San Pedro Domes geothermal field and supports continued exploration through detailed geophysical, geochemical, and drilling investigations. Overall, this study demonstrates the value of integrating indirect geophysical methods for early–stage geothermal assessment in volcanic regions of the Trans–Mexican Volcanic Belt.

Key words: Geothermal exploration, Hydrothermal system, Trans–Mexican Volcanic Belt, Potential geophysical data, Amado Nervo Volcano.

Resumen

Este estudio presenta un modelo conceptual del sistema hidrotermal asociado al Volcán Amado Nervo, en el occidente de México, mediante la integración de información geológica, gravimétrica, magnética y geotérmica. Los objetivos fueron caracterizar la estructura subsuperficial del sistema volcánico, identificar firmas geofísicas potencialmente asociadas a la alteración hidrotermal y proporcionar una estimación preliminar del potencial geotérmico explotable de la región. Los datos aeromagnéticos fueron procesados mediante técnicas de reducción al polo, señal analítica, continuación ascendente y análisis espectral, mientras que los datos gravimétricos se derivaron del Modelo Gravimétrico Global (GGM+) y se transformaron en anomalías de Bouguer simple y residual. El método del espectro de potencia promediado radialmente se aplicó para estimar la profundidad de las fuentes magnéticas, la profundidad del punto de Curie, el gradiente geotérmico y el flujo de calor conductivo. Los resultados geofísicos revelan una asociación entre zonas de atenuación magnética, mínimos de señal analítica y anomalías gravimétricas negativas, delimitando dominios estructuralmente complejos caracterizados por una reducción de la densidad y la magnetización. El análisis espectral permitió estimar una profundidad del punto de Curie de aproximadamente 7.5 km, a partir de la cual se calcularon un gradiente geotérmico de 70.8 °C·km⁻¹ y un flujo de calor conductivo de 176 mW·m⁻². Estos valores son consistentes con los reportados para regiones geotérmicas del Cinturón Volcánico Trans–Mexicano. La integración de la información geofísica y geológica permitió desarrollar un modelo conceptual hidrotermal correspondiente a un sistema volcánico–hidrotermal estructuralmente controlado y potencialmente sustentado por una fuente de calor intrusiva residual. La evaluación preliminar del recurso geotérmico se realizó mediante el método volumétrico Heat in Place acoplado con 9,999 simulaciones de Monte Carlo. Los resultados arrojaron un potencial geotérmico promedio de 19.75 MW, con valores simulados que oscilan entre 5 y 34 MW. Aunque esta estimación está sujeta a incertidumbres relacionadas con la geometría del reservorio y los parámetros termofísicos empleados, el recurso estimado sitúa al sistema volcánico Amado Nervo dentro del rango de prospectos geotérmicos de escala media en México. La arquitectura hidrotermal propuesta presenta similitudes con la del campo geotérmico cercano de Domes de San Pedro y respalda la continuidad de las actividades de exploración mediante investigaciones geofísicas, geoquímicas y de perforación más detalladas. En conjunto, este estudio demuestra el valor de integrar métodos geofísicos indirectos para la evaluación geotérmica en etapas tempranas dentro de regiones volcánicas del Cinturón Volcánico Trans–Mexicano.

Palabras clave: Exploración geotérmica, Sistema hidrotermal, Cinturón Volcánico Trans–Mexicano, Datos geofísicos potenciales, Volcán Amado Nervo.

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1. Introduction

The Trans–Mexican Volcanic Belt (TMVB) is one of the regions with the highest geothermal potential in Mexico, owing to tectonic activity driven by the subduction of the Rivera and Cocos Plates beneath the North American Plate (Figure 1). This geodynamic configuration favors the development of hydrothermal systems dominated by fluid convection, in which the primary heat source originates from shallow magmatic bodies (Prol–Ledezma and Morán–Zenteno, 2019). Consequently, the TMVB hosts multiple geothermal exploration projects, several of which are currently under commercial operation.

Within this volcanic province, the state of Nayarit contains several volcanic edifices of geothermal significance, including Sangangüey, Ceboruco, San Juan, Tepetiltic, Las Navajas, the San Pedro Domes, and the Amado Nervo Volcano. A key structural feature in the region is the San Pedro–Ceboruco Graben (SPCG), an east–west–trending tectonic depression formed by normal faulting associated with crustal extension in the TMVB. The graben exhibits evidence of recent tectonic activity and Holocene volcanism, conditions that are favorable for the development of active hydrothermal systems (Ferrari *et al.*, 2003). Its central sector hosts the San Pedro Domes geothermal field, with an installed capacity of 35.5 MW, recognized as the first geothermal power plant operated by a private company in Mexico (Reyes–Orozco *et al.*, 2019).

Scientific research in the SPCG has been extensive yet heterogeneous. Several decades ago, the Comisión Federal de Electricidad conducted geological, geophysical, and geochemical surveys, the results of which remain in confidential internal reports. Over the past 30 years, numerous studies have contributed to understanding the region's geology, stratigraphy, and eruptive history (Ferrari *et al.*, 1999, 2003; Petrone *et al.*, 2001; Sieron and Siebe, 2008). Concurrently, a wide range of geophysical investigations, including gravity, magnetic, seismic, and electromagnetic studies, have provided insights into the subsurface structure of the graben (Venegas, 1995; Sánchez *et al.*, 2009; Rodríguez–Uribe *et al.*, 2013; Fernández–Córdoba *et al.*, 2017; Sawires and Aboud, 2019; Núñez–Cornú *et al.*, 2020; Fuentes–Arreazola *et al.*, 2021; Muñoz–Burbano *et al.*, 2024). Additionally, several studies have proposed models of the hydrothermal system hosted within the graben, primarily based on geological information and well data (Reyes–Orozco *et al.*, 2019; Rodríguez *et al.*, 2019).

At the southeastern end of the SPCG lies the Amado Nervo Volcano, a shield–type edifice composed mainly of basaltic lava flows, located in a structural context similar to that of the San Pedro Domes geothermal field. Despite these favorable conditions, this volcano has received comparatively little scientific attention,

with most studies focusing on general geological descriptions (e.g., Ferrari *et al.*, 2003). The absence of integrated geophysical studies represents a significant gap in understanding its internal structure and geothermal potential.

To address this gap, the present study integrates aeromagnetic and gravity data with existing geological and hydrogeological information to characterize the geophysical configuration of the Amado Nervo Volcano. Through the analysis and interpretation of potential–field anomalies, we identify key subsurface properties, delineate zones potentially associated with hydrothermal alteration, and propose a preliminary conceptual model and geothermal potential of the hydrothermal system that may be hosted beneath the volcanic edifice. This work represents the first detailed geophysical characterization of the area and provides a robust foundation for subsequent stages of advanced geothermal exploration.

2. Study area

2.1. Tectonic–geological setting and eruptive history

The Amado Nervo Volcano is a Quaternary shield volcano (0.22 Ma) located in the state of Nayarit (21.150399°N, –104.756680°W; 1447 m.a.s.l.), near the northwestern limit of the TMVB. It lies within the SPCG, a tectonic structure associated with the Tepic–Zacoalco Rift, which marks the boundary between the Jalisco Block and the North American Plate (Figure 1). Together with the Chapala and Colima rifts, the Tepic–Zacoalco Rift defines a triple–junction zone near the city of Guadalajara (Nelson, 1980; Stock, 1993; Ferrari *et al.*, 2003). This tectonic framework is characterized by crustal extension, normal faulting, and recent volcanism.

The geological evolution of the SPCG spans from the late Miocene to historical times and is marked by shifts in magma composition and tectonic activity (Figure 2). Intense rhyolitic volcanism dominated the early Pliocene (5–4.2 Ma), followed by mafic eruptions associated with regional extension. Around 1.1 Ma, activity evolved into a more differentiated bimodal regime, leading to the formation of the San Pedro–Cerro Grande volcanic complex, including dacitic–andesitic domes and the Tepetiltic stratovolcano (~480 ka) (Ferrari *et al.*, 1999, 2003; Petrone *et al.*, 2001).

A major event in this sequence was the collapse of the San Pedro caldera between ~450 and 283 ka, followed by intracaldera dome growth (e.g., Los Ixletes, Los Ocotes, Cerro Lobos) and the Cuastecomate Plinian eruption (~210 ± 260 ka), which produced widespread pyroclastic deposits. Subsequently, the Amado Nervo Volcano developed (220 ± 29 ka) through the

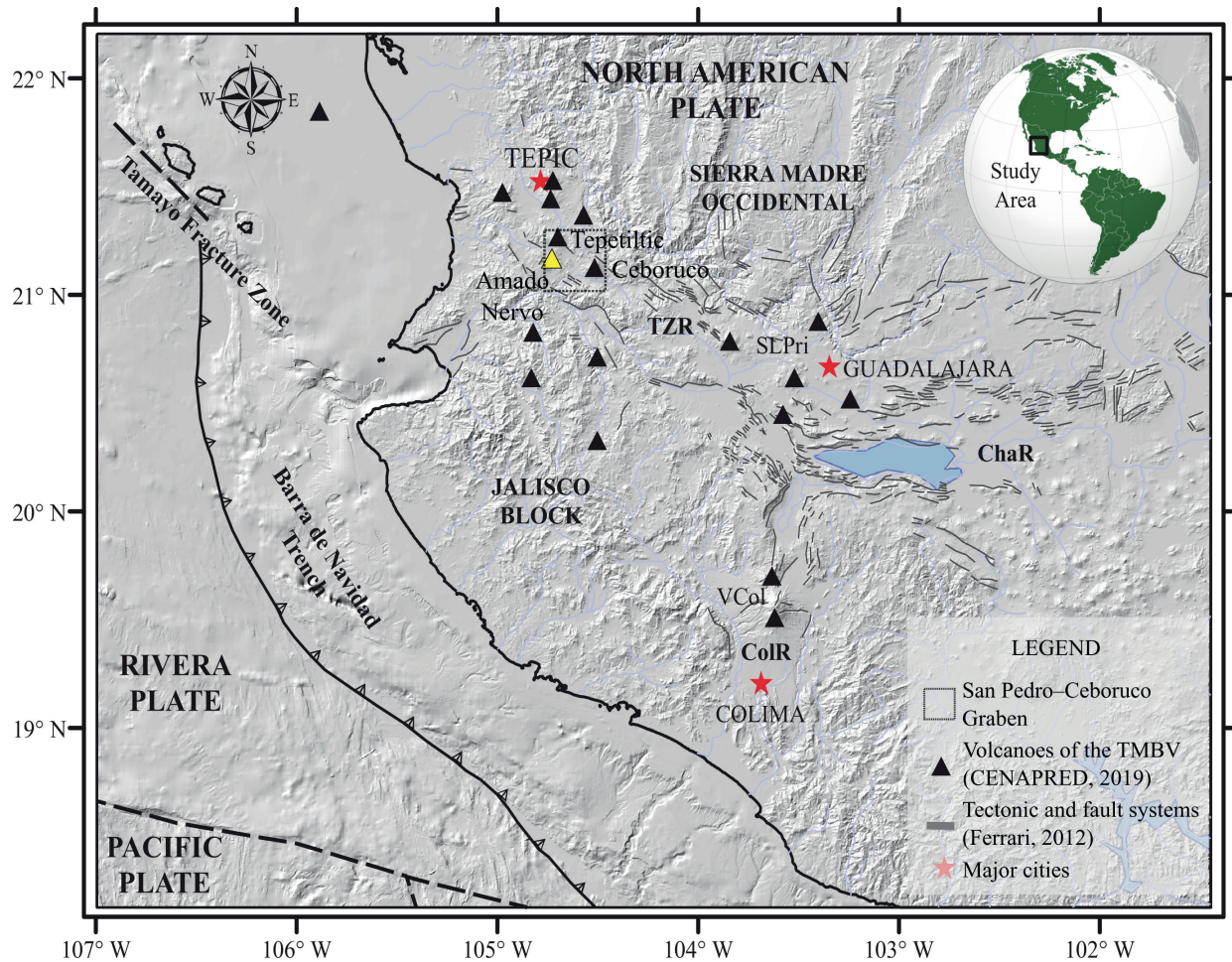


Figure 1. Location map of the Amado Nervo Volcano illustrating its spatial relationship with major tectonic domains and plate boundaries. The Tepic–Zacoalco (TZR), Chapala (ChaR), and Colima (ColR) rift zones are also indicated. Inset: Location of the study area within the American continent. Modified from Fuentes–Arreazola *et al.* (2021).

effusion of mildly alkaline basaltic lavas, representing a late effusive stage that contrasts with the predominantly silicic activity of the San Pedro complex (41 ± 20 ka) and suggests magmatic rejuvenation during the Quaternary. The southeastern sector hosts the Ceboruco Volcano, the youngest and most active center ($45\text{--}0$ ka), with historical eruptions [A.D. 930, 1870–1875] (Núñez *et al.*, 2022) and persistent fumarolic activity indicating an active hydrothermal system (Ferrari *et al.*, 2003; Sieron and Siebe, 2008; Núñez–Cornú *et al.*, 2020).

At a more local scale, the Amado Nervo Volcano is situated within the San Pedro–Amado Nervo subgraben, limited by NW–SE and NE–SW–oriented normal faults and by minor strike–slip faults in its southeastern portion (Figure 2). Major structures in this area include the Borbollón, Milpillas–Cerro Grande, and Las Cebollas faults, as well as the Amado Nervo strike–slip fault. This structural configuration strongly influences regional morphology (Reyes–Orozco *et al.*, 2019) and, speculatively, also the fluid circulation.

According to the geological map (Figure 2), constructed using data from INEGI (2025) and the Mexican Geological Survey (Servicio Geológico Mexicano, 2025), the study area comprises the following surface lithologies: rhyolitic ignimbrite, pumice–fall deposits, alluvium, Tertiary rhyolitic breccia, Quaternary basaltic lava flow, and Quaternary andesitic lava flow. Locally, sandstone–limestone units may also crop out, although they are more commonly present at depths greater than 400 m. At deeper stratigraphic levels, rhyolitic ignimbrites, calcarenite, Tertiary andesitic lava flows, and granitic bodies are identified. Basaltic breccia is exposed at the eruptive vent of the Amado Nervo Volcano, consistent with its effusive origin.

2.2. Hydrogeological and hydrothermal setting

From a hydrological perspective, the region exhibits a predominantly dendritic drainage pattern, transitioning to a radial configuration over the volcanic edifices. Surface runoff

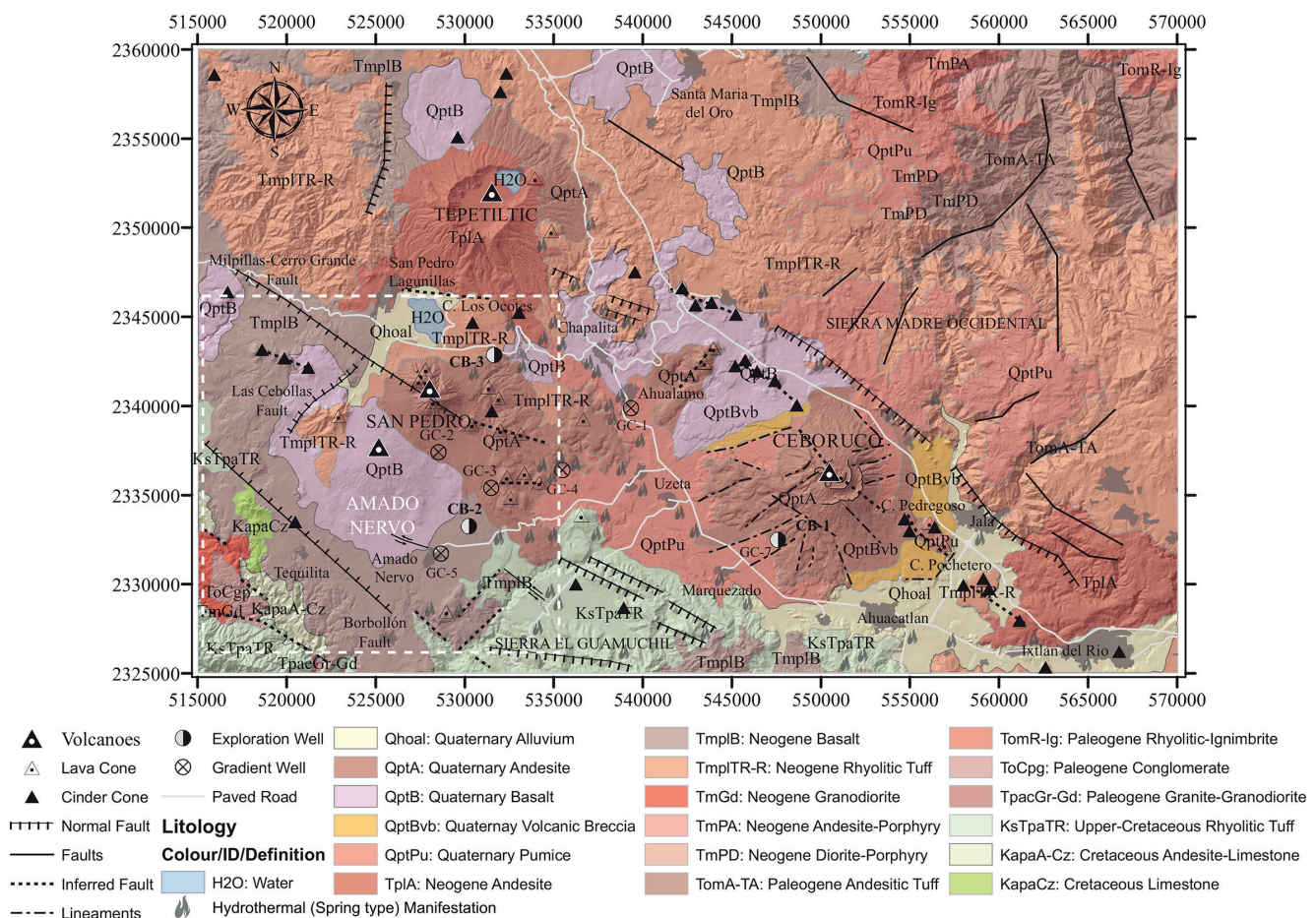


Figure 2. Regional to local geological map of the San Pedro–Ceboruco Graben. Exploration and gradient wells, as well as lava and scoria cones, are shown following Ferrari *et al.* (2003). Fault traces and lithological units are compiled from Ferrari *et al.* (2012), SGM (2025), and INEGI (2025). Hydrothermal manifestations location from Tello–Hinojosa (1994). Inset: The dotted white line delineates the specific area analyzed using geophysical data near the San Pedro–Amado Nervo subgraben.

and stream channels generally flow toward the southwest. The volcanic units exposed in the study area may exhibit variable porosity and permeability depending on their lithology, degree of fracturing, and volcanic textures. Although quantitative hydrogeological data are currently unavailable for these units, highly fractured zones associated with regional and local structures may plausibly act as preferential pathways for infiltration, groundwater recharge, and fluid circulation within shallow aquifers and potential hydrothermal systems.

Three unconfined aquifers have been identified in the vicinity of Amado Nervo Volcano: Valle de Compostela, Valle de Banderas, and Ixtlán–Ahuacatlán. Although detailed hydrogeological properties of these systems remain poorly constrained, groundwater occurrence and flow are likely influenced by lithological heterogeneity and rock fracturing. Available piezometric information is limited and should be regarded only as a regional hydrogeological reference. For example, groundwater-level measurements obtained during a well-leveling survey conducted in

October 2017 indicate that the water table in alluvial deposits near the community of Juan Escutia, approximately 10 km west of the volcano, ranged between 10 and 30 m below ground surface (Comisión Nacional del Agua, 2025). The referenced report also indicates that slight groundwater-level declines typically begin in November due to agricultural groundwater extraction following the rainy season. Because these measurements correspond to a nearby alluvial aquifer rather than the volcanic edifice itself, they are presented solely to provide a general regional context regarding groundwater occurrence and flow conditions.

The geochemical study of hot springs surrounding the Ceboruco Volcano (Figure 2), conducted by Tello–Hinojosa (1994), revealed an active hydrothermal system linked to both the volcanic edifice and its immediate structural framework. The hydrothermal manifestations exhibit chemical compositions consistent with the heating of meteoric waters that infiltrate through highly permeable zones, primarily fractures and faults associated with the regional geological–structural configuration.

These fluids then recirculate toward the surface, driven by elevated thermal gradients.

Within this regional setting, geothermal gradient measurements from nearby exploratory wells indicate significant thermal anomalies around the Amado Nervo Volcano (Figure 2): GC–5 (~6.9 km south–southeast) with $53.2\text{ }^{\circ}\text{C km}^{-1}$; CB–2 (~6.1 km southeast) with $102.5\text{ }^{\circ}\text{C km}^{-1}$; GC–3 (~7 km east–southeast) with $103.6\text{ }^{\circ}\text{C km}^{-1}$, and CB–3 (~7.8 km northeast) with $52.4\text{ }^{\circ}\text{C km}^{-1}$ (Espinoza–Ojeda *et al.*, 2023). These values support the presence of a high–temperature regime consistent with the development of deep hydrothermal systems.

Additionally, the geochemical signatures, including characteristic concentrations of silica, bicarbonate, and other major ions, suggest a medium– to high–temperature reservoir influenced by water–rock interaction processes. The spatial distribution of springs and their alignment with major structural lineaments further indicate that hydrothermal flow is channeled along faults that serve as preferential pathways for upward fluid migration.

3. Data and method of analysis

3.1. Geophysical data

The aeromagnetic data used in this study were obtained from a 1983 survey conducted by the Mexican Geological Survey (SGM, 2005). The survey was carried out using GPS navigation along flight lines spaced 1,000 m apart, at an average elevation of 300 m above ground level. The total magnetic field was computed after applying corrections for aircraft motion, diurnal variation, and flight–line leveling. The residual magnetic field (RMF) was derived by subtracting the International Geomagnetic Reference Field (1980), using reference values of total intensity (43,415 nT), inclination ($47^{\circ}49'$), and declination ($8^{\circ}52'$). The SGM produced a 1:50,000–scale contour map of the RMF with 10 nT contour intervals. For this study, a digital ASCII grid with 200 m spacing and an extent of $20 \times 20\text{ km}$ (400 km^2) was acquired from the SGM and centered near the Amado Nervo Volcano. This grid served as the basis for aeromagnetic processing and analysis.

In this study, gravity data from the Global Gravity Model GGM+ (Hirt *et al.*, 2013) were used. This model is a modern tool for regional gravity analysis, integrating satellite measurements from missions such as GRACE and GOCE. GGM+ provides an approximate spatial resolution of 200 m, comparable to the station spacing typically employed in terrestrial gravity surveys for local characterization of volcanic structures. Binary files corresponding to the N20W105 cell were downloaded, containing gravity acceleration values (N20W105.ga) and gravity disturbance

values (N20W105.dg). From these files, data within the study area were extracted, yielding a grid of 90 rows by 97 columns (8,730 nodes) that served as the basis for regional analysis and geophysical interpretation.

3.2. Aeromagnetic and gravimetric data processing and analysis

The residual magnetic field data were processed through a series of standard transformations, including reduction to the pole (RTP), analytic signal amplitude, and upward continuation, to enhance magnetic source signatures and facilitate geological interpretation. Following the RTP transformation, the workflow described by Fuentes–Arreazola *et al.* (2023) and references therein was applied to compute and correct the radially averaged power spectrum. This spectral analysis enabled estimation of the upper and lower depths of the magnetic sources responsible for the observed anomalies. The lower bound of these sources was interpreted as the Curie Point Depth, from which the thermal gradient and conductive heat flow were subsequently estimated following the formulations of Turcotte and Schubert (2002).

The gravity disturbance data derived from the GGM+ model correspond to Free–Air Anomaly values, defined as the difference between the measured gravity at a specific point on Earth's surface and the theoretical gravity, after correction for elevation and the atmospheric mass between the terrain surface and the satellite (Seigel, 1995; Hildenbrand *et al.*, 2002). From these disturbance values, the observed gravity at each hypothetical station was reconstructed. Latitude, elevation, and Bouguer slab corrections were then applied to compute the simple Bouguer anomaly (Telford *et al.*, 2012; Seigel, 1995). To isolate the local residual gravity field, a band–pass filter with a long–wavelength cutoff of 20 km, corresponding to the lateral extent of the analyzed grid, was applied to remove regional trends.

From the RTP magnetic anomaly upward continued to 1 km, together with the residual simple Bouguer anomaly, a 10–km–extended west–east profile was extracted across the study area. This profile was subsequently modeled to approximate the regional geological framework. All aeromagnetic and gravimetric processing steps, including data reduction, spectral analysis, filtering, and potential–field transformations, were carried out using Oasis montaj software (Seequent Ltd., 2025).

3.3. Geothermal potential assessment

The volumetric Heat in Place method is commonly used in preliminary geothermal assessments to estimate the thermal energy stored within a volume of hot rock. The recoverable thermal energy [Q_{rec}] can be expressed as (Equation 1):

$$Q_{rec} = R_g V [\phi \rho_w c_w + (1 - \phi) \rho_r c_r] (T_r - T_{ref}) \quad \text{Eq.1}$$

where, V is the reservoir volume, ϕ is the porosity, ρ is the water and rock densities, c is the water and rock specific heat, T_r is the reservoir temperature, T_{ref} is the reference temperature, and R_g is the recovery factor.

To account for parameter uncertainty, Monte Carlo simulations were performed using a custom MATLAB script developed for this study following the methodology described by Muffler and Cataldi (1978), Hiriart-Le Bert *et al.* (2011), and Garg and Combs (2015). The simulations generated probabilistic estimates of geothermal potential, and the distribution's mean was adopted as the most likely resource estimate.

4. Results

4.1. Gravimetric and magnetic anomalies

The geophysical results are presented beginning with the gravimetric anomalies, followed by the magnetic anomalies and their transformations. Together, these observations provide the basis for developing the hydrothermal conceptual model and for assessing the geothermal potential of the study area.

The simple Bouguer anomaly ranges from -81.74 mGal to -149.13 mGal and exhibits a well-defined regional trend across the study area (Figure 3a). Higher values (~ -90 mGal) occur in the southwestern sector, whereas lower values (~ -120 mGal) are concentrated toward the northeast. Subtle gravity gradients are observed throughout the area, several of which are spatially coincident with major northwest-southeast structural features, including the Borbollón and Milpillas-Cerro Grande faults.

At the local scale, the Amado Nervo Volcano (ANV) is located within a domain characterized by intermediate Bouguer anomaly values of approximately -116 mGal, whereas Cerro Las Tetillas coincides with a relative gravity minimum of approximately -129 mGal. The San Pedro Domes are situated within a well-defined circular anomaly that reaches minimum values of approximately -140 mGal. The geothermal wells of the San Pedro Domes project are located in the northeastern sector of this anomaly, within an area characterized by negative Bouguer anomaly values relative to the surrounding region.

To isolate the local gravity response, a band-pass filter with a long-wavelength cutoff of 20 km, corresponding to the lateral extent of the data block, was applied. The resulting residual Bouguer anomaly (Figure 3b) ranges from -29.62 mGal to 15.53 mGal, with a mean value of 0.02 mGal. After removing the regional trend, a continuous negative anomaly is observed

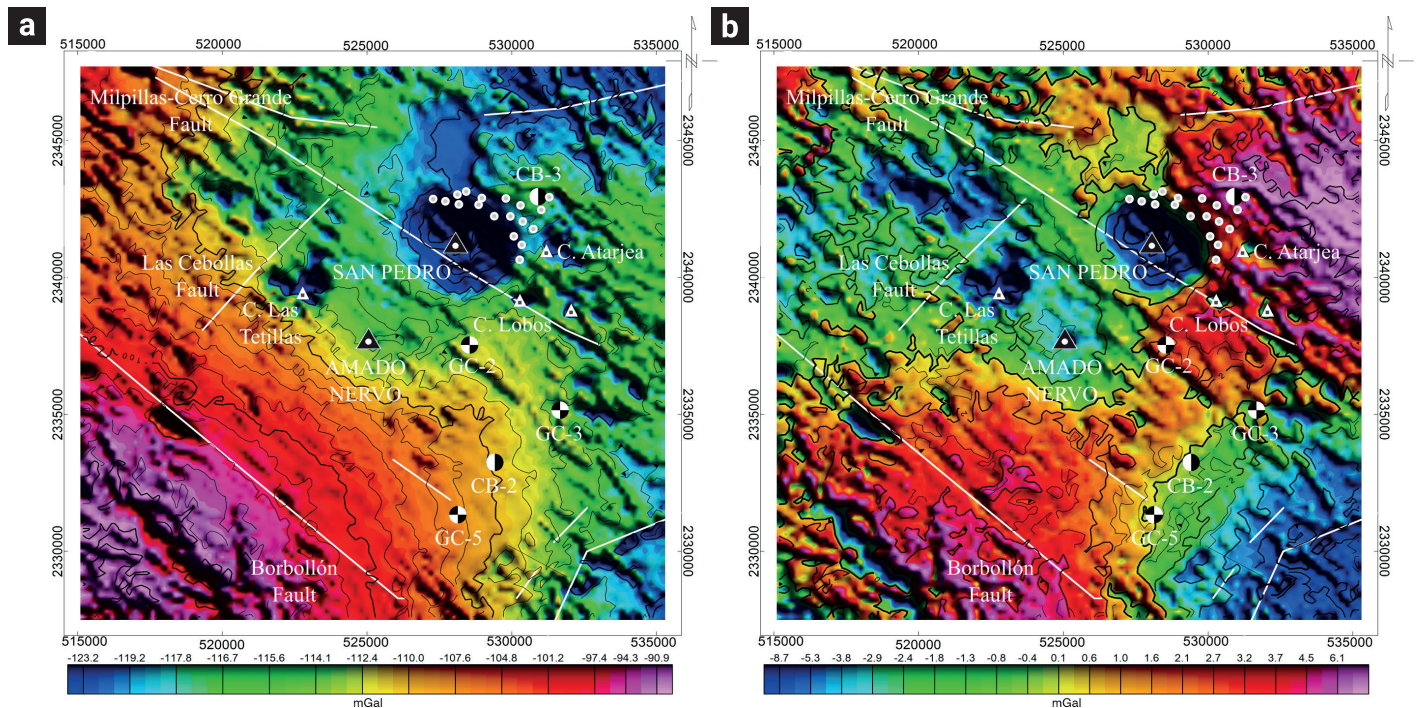


Figure 3. (a) Simple Bouguer and (b) residual/local Bouguer anomaly map of the Amado Nervo Volcano and surrounding areas. Colors and contour lines represent the distribution of gravity anomaly intensities. White lines correspond to geological faults compiled from INEGI (2025) and SGM (2025). Triangular symbols indicate volcanic structures (Ferrari *et al.*, 2003). Circles with alternating fill mark the locations of geothermal exploration wells (CB-2 and CB-3) and gradient wells (GC-2, GC-3, and GC-5) reported by Ferrari *et al.* (2003). Solid white-gray circles denote the wells of the Doms de San Pedro geothermal project.

beneath both the Amado Nervo Volcano and Cerro Las Tetillas. The residual anomaly reaches -4.32 mGal at the ANV and attains its lowest value at Cerro Las Tetillas (-17.83 mGal). This negative domain is located between the Las Cebollas fault to the northwest and the Milpillas–Cerro Grande fault to the northeast.

Higher residual values are observed south, east, and northeast of the Amado Nervo Volcano. In contrast, the San Pedro Domes coincide with a well-defined negative residual anomaly, with values below -20 mGal.

Figure 4a displays the residual magnetic field anomaly of the Amado Nervo Volcano and surrounding areas, with values ranging from -337.20 to $1,120.86$ nT. Positive anomalies occur primarily over the central sector and toward the south, coinciding with the Amado Nervo Volcano edifice and the San Pedro Domes. An additional positive domain is observed to the northeast in the vicinity of Tepetitlic Volcano, whereas negative anomalies are distributed along the western, northern, and eastern margins of the Amado Nervo Volcano. A well-defined dipolar anomaly is observed at the volcanic vent, characterized by magnetic values ranging from approximately -157 nT to 630 nT. The boundaries of the principal anomaly are spatially coincident with the Borbollón, Las Cebollas, and Milpillas–Cerro Grande faults systems.

To remove the effects of magnetic field declination and inclination, the total magnetic field data were transformed to the

magnetic pole (Figure 4b). The resulting anomaly ranges from -485.46 nT to $1,391.61$ nT and exhibits a predominant northwest–southeast orientation. The principal anomaly remains centered on the Amado Nervo Volcano and comprises three high-intensity lobes with amplitudes of $1,384$ nT, $1,061$ nT, and 976 nT together with a fourth lower-intensity lobe of 135 nT. The anomaly boundaries are spatially coincident with major regional faults.

To better isolate the magnetic sources within the study area, the analytical signal of the reduced-to-the-pole magnetic anomaly was calculated (Figure 5a). The Amado Nervo Volcano exhibits analytical signal values of approximately 1.30 nT·m⁻¹, whereas the San Pedro Domes are characterized by values of approximately 0.50 nT·m⁻¹. Other volcanic structures also exhibit notable analytical signal amplitudes: Cerro Tetillas, located northwest of the Amado Nervo Volcano, exhibits a value of 0.83 nT·m⁻¹, whereas Cerro Lobos, situated southeast of the San Pedro Domes, reaches 1.05 nT·m⁻¹ and is spatially aligned with the Milpillas–Cerro Grande fault. Three of the five geothermal gradient wells are located in zones with low analytical signal values ranging from 0.07 to 0.15 nT·m⁻¹. Similarly, wells from the San Pedro Domes geothermal project occur in areas with analytical signal values between 0.03 and 0.30 nT·m⁻¹.

To obtain a regional perspective of the magnetic field, a 1 km upward continuation was applied to the reduced-to-

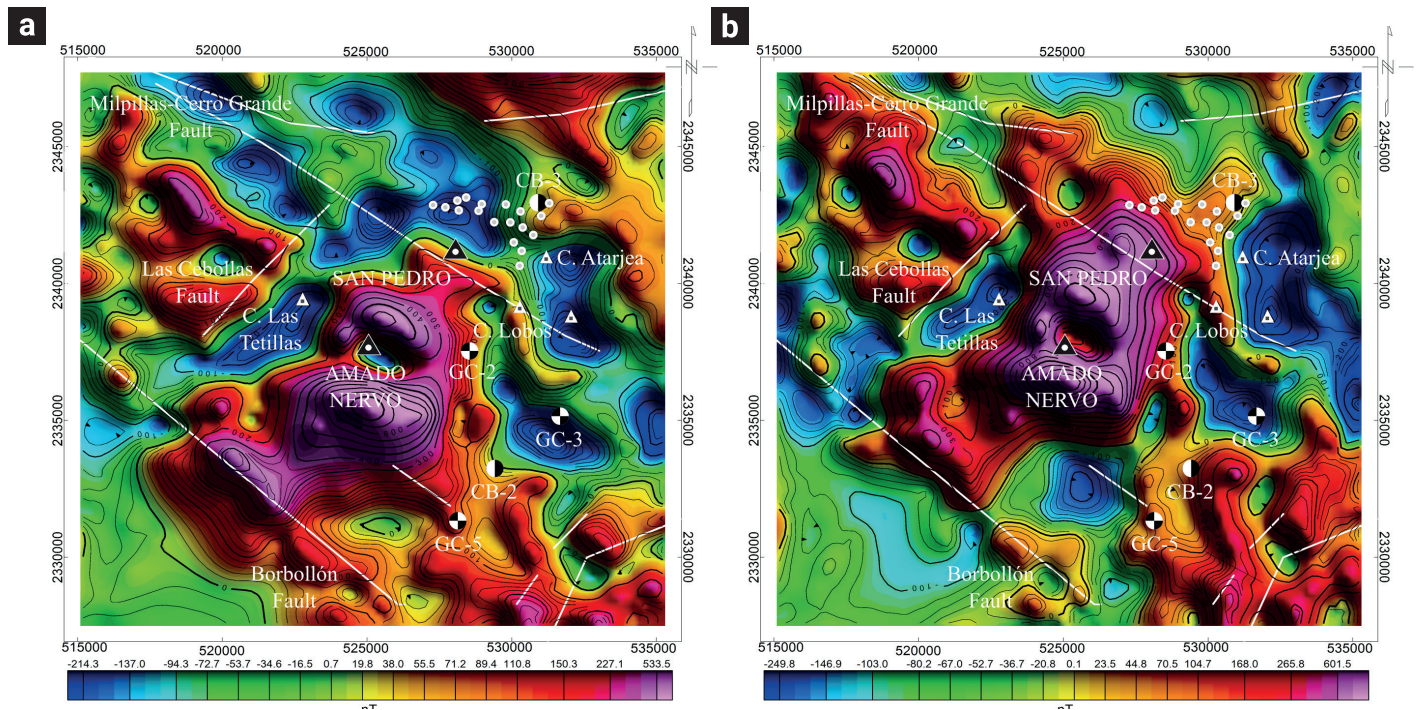


Figure 4. (a) Residual magnetic field and (b) magnetic field reduced to the magnetic pole anomaly map of the Amado Nervo Volcano and surrounding areas. Colors and contour lines represent the distribution of magnetic field intensities.

the-pole anomaly (Figure 5b). The resulting anomaly ranges from -157 to 580 nT. Following continuation, the principal magnetic anomaly remains prominent, with amplitudes approaching 500 nT, and remains spatially associated with the Amado Nervo Volcano.

4.2. Hydrothermal conceptual model

The hydrothermal conceptual model for the Amado Nervo volcanic system was developed by integrating geophysical, geothermal, and geological information (Figure 6). Spectral analysis of the reduced-to-the-pole magnetic anomaly was used to constrain the depth distribution of magnetic sources and estimate the lower magnetic boundary. Together with gravimetric anomalies, magnetic anomaly patterns, mapped structural features, and available geothermal data, these results provide the framework for the proposed model.

Magnetic source depths were estimated from a radially averaged power spectrum calculated over a 20×20 km block of the reduced-to-the-pole anomaly. Using the spectral slope analysis of Ravat *et al.* (2007) and Fedi *et al.* (1997), the centroid depth was estimated at $Z_0 = -4.34$ km and the top of the magnetic sources at $Z_t = -1.18$ km. Applying the methodology of Okubo *et al.* (1985) and Tanaka *et al.* (1999), the lower bound of the magnetic source was estimated at $Z_B = -7.50$ km.

Assuming Z_B corresponds to the Curie point depth and using a Curie temperature of approximately 580°C , a geothermal gradient of $70.79^\circ\text{C}\cdot\text{km}^{-1}$ was calculated following Núñez–Demarco *et al.* (2020), using a surface temperature of 28°C and a flight altitude of 0.3 km. Using a thermal conductivity of $2.5\text{ W}\cdot\text{m}^{-1}\cdot^\circ\text{C}^{-1}$ (Turcotte and Schubert, 2002), a conductive heat flow of approximately $176\text{ mW}\cdot\text{m}^{-2}$ was obtained. Based on the estimated geothermal gradient, subsurface temperatures were calculated as a function of depth, yielding approximately 150°C at 1.5 – 1.7 km depth and exceeding 700°C at approximately 11 km depth.

The conceptual model comprises a shallow volcanic sequence overlying deeper basement rocks and intrusive units. A transition in geophysical properties is identified near sea level, corresponding to the boundary between volcanic and deeper lithological domains. Geothermal gradient wells, magnetic anomalies, gravity anomalies, and mapped structural features were integrated to define the principal components of the proposed hydrothermal system.

4.3. Geothermal potential estimation

The conceptual model proposes a potential geothermal reservoir located beneath the inferred cap-rock sequence and hosted within fractured crystalline basement rocks intruded

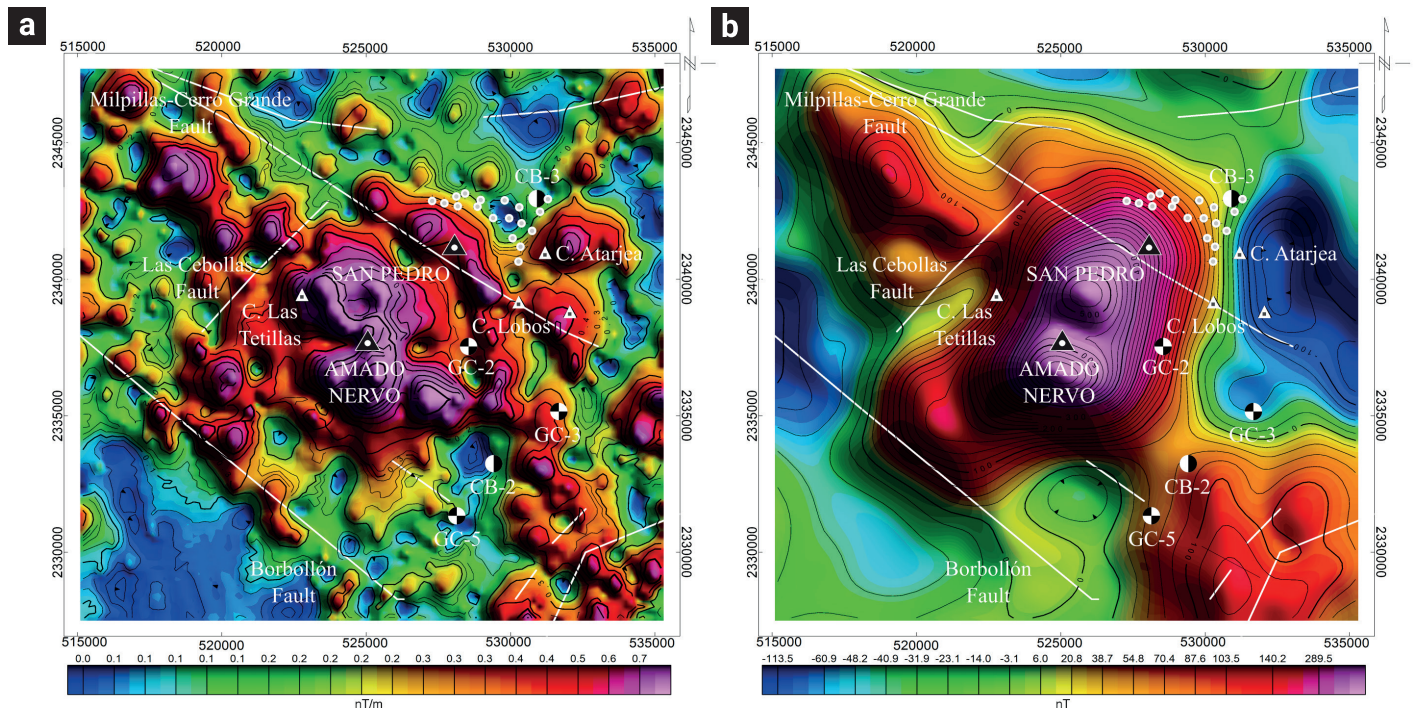


Figure 5. (a) Analytical signal of the residual magnetic field reduced to the magnetic pole, and (b) upward-continued (1 km) anomaly map of the Amado Nervo Volcano and surrounding areas.

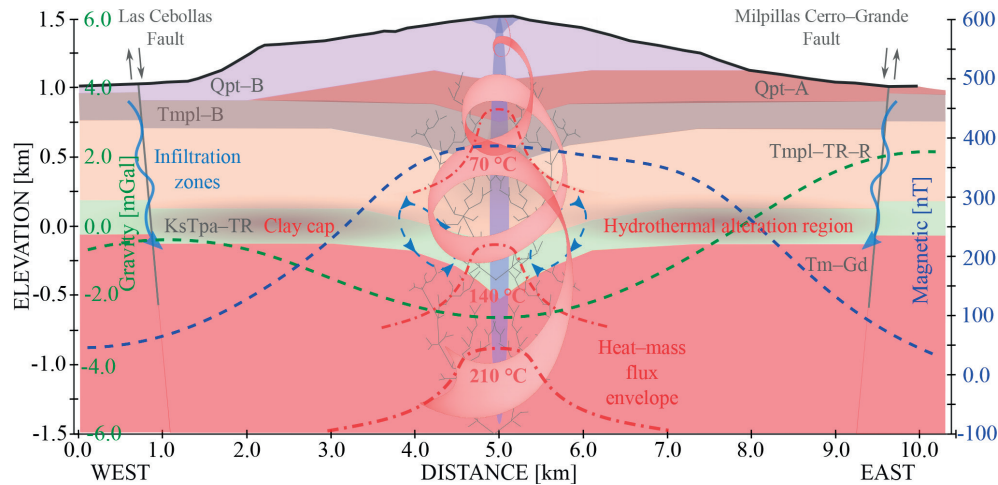


Figure 6. Conceptual hydrothermal model of the Amado Nervo volcanic system. The inset illustrates the residual gravity anomaly and the 1-km upward-continued magnetic anomaly along a 10-km west–east profile. Geological units correspond to those shown in Figure 2. Regional fault systems are incorporated as structural features associated with fluid recharge, permeability enhancement, and hydrothermal fluid migration.

by intermediate–composition bodies. The model identifies this domain as a potential target for future exploratory drilling.

The integration of magnetic, gravity, and geothermal–gradient data reveals a spatial association between negative magnetic anomalies, analytical signal attenuation, negative gravity anomalies, and elevated geothermal gradients. These features define geophysical domains characterized by contrasting physical properties within the study area.

The proposed conceptual model exhibits several similarities to the San Pedro Domes geothermal field. In both areas, hydrothermal alteration zones occur at depths of approximately 1 km, temperatures near 150 °C have been reported or estimated, and

crystalline basement rocks of the Jalisco Block occur at depths of approximately 1.2–1.8 km beneath volcanic sequences (Reyes–Orozco *et al.*, 2019).

A set of key parameters derived from geological, geophysical, and geothermal data was integrated into a volumetric Heat in Place model coupled with Monte Carlo simulations (Table 1). Based on a MATLAB implementation developed for this study and 9,999 simulations, a mean geothermal potential of 19.75 MW with an uncertainty of ± 14.41 MW was estimated. The resulting distribution spans approximately 5 MW under conservative scenarios to approximately 34 MW under more favorable conditions.

Table 1. Range of values for the variables involved in the volumetric model used to estimate geothermal potential.

Variables for the Volumetric (Heat In Place) Model		
Parameter	Minimum value	Maximum value
Project lifetime [years]	25	35
Separation pressure [bar]	5	12
Reservoir area [km ²]	1	15
Reservoir thickness [km]	0.10	1.50
Reservoir temperature [°C]	140	250
Ambient temperature [°C]	20	35
Reservoir rock porosity [%]	0.01	0.17
Reservoir rock density [kg·m ⁻³]	2400	2900
Reservoir rock specific heat [kJ kg ⁻³ ·°C ⁻¹]	0.60	1.20
Reservoir recovery factor [%]	0.05	0.25
Cycle efficiency [%]	0.10	0.25

5. Discussion

5.1. Geophysical interpretation of gravity and magnetic anomalies

The gravimetric anomalies reveal significant spatial variations that may reflect changes in subsurface density and structural conditions across the Amado Nervo volcanic complex. At the regional scale, the southwest–northeast decrease in Bouguer anomaly values suggests the presence of large-scale density contrasts, while the spatial coincidence between gravity gradients and the Borbollón and Milpillás–Cerro Grande fault systems indicates that regional structures may influence the distribution of these contrasts. Although the observed gradients are subtle, they could be associated with lithological boundaries, zones of enhanced fracturing, or other structurally controlled heterogeneities within the volcanic system.

Local gravity anomalies further highlight variations in the physical properties of the volcanic centers. The intermediate Bouguer anomaly values associated with the Amado Nervo Volcano contrast with the pronounced gravity minima observed at Cerro Las Tetillas and the San Pedro Domes. These negative anomalies may reflect reduced bulk density resulting from enhanced fracturing, hydrothermal alteration, volcanic deposits with elevated porosity, structural collapse, or a combination of these processes. The continuous negative residual anomaly extending beneath both the Amado Nervo Volcano and Cerro Las Tetillas suggests that these volcanic centers may be linked by a common subsurface domain characterized by reduced density. Likewise, the distribution of residual anomalies around the volcanic complex, including relatively high values south, east, and northeast of the Amado Nervo Volcano, points to significant lateral heterogeneity in subsurface properties. These higher residual values may correspond to denser lithological units or structurally coherent blocks that have experienced comparatively limited alteration.

The geometry and spatial distribution of the gravity anomalies further emphasize the importance of tectonic controls. The negative residual anomaly shared by the Amado Nervo Volcano and Cerro Las Tetillas is located between the Las Cebollas and Milpillás–Cerro Grande faults, suggesting that regional fault systems may influence the distribution of density contrasts and the organization of subsurface structures. More broadly, the occurrence of the Amado Nervo Volcano, Cerro Las Tetillas, and the San Pedro Domes within a continuous negative gravimetric domain is consistent with a regionally extensive zone of reduced density and structural complexity. This interpretation agrees with the structural model of Reyes–Orozco *et al.* (2019), which describes a southeastward–deepening faulted block system beneath

the San Pedro Domes area, and with the tectonic framework proposed by Venegas (1995), in which the San Pedro–Tepetitlic Block represents the most subsided structural domain in the region. Such structural depressions may constitute long-lived zones of crustal weakness that favor magma ascent and volcanic activity within the San Pedro–Ceboruco Graben.

Taken together, the gravity data suggest that the Amado Nervo volcanic complex is characterized by pronounced density heterogeneity controlled by both volcanic and tectonic processes. Although the origin of individual anomalies cannot be uniquely determined from gravity data alone, the regional and residual anomaly patterns are consistent with a structurally complex volcanic system shaped by fault-controlled deformation, volcanic construction, and density-reducing processes such as fracturing and alteration.

The magnetic anomalies provide complementary constraints on the structural and magmatic framework of the Amado Nervo volcanic system. At the regional scale, the residual and reduced-to-the-pole magnetic anomaly maps reveal a heterogeneous magnetic fabric characterized by positive magnetic domains associated with the Amado Nervo Volcano, the San Pedro Domes, and other volcanic centers, as well as magnetic lows distributed along the margins of the volcanic complex. The spatial correspondence between several magnetic gradients and major fault systems suggests that tectonic structures may have significantly influenced the distribution of magnetic contrasts and the organization of the volcanic edifice.

The prominent dipolar anomaly centered on the Amado Nervo Volcano suggests the presence of a strongly magnetized source beneath the volcanic edifice. Similar magnetic signatures are commonly associated with basaltic volcanic rocks, shallow intrusive bodies, or volcanic units that preserve significant remanent magnetization (Blakely, 2009; Telford *et al.*, 2012). Furthermore, the reduced-to-the-pole anomaly map reveals a series of northwest–southeast–trending magnetic domains whose boundaries closely coincide with the principal regional faults. The clustering of high-amplitude magnetic lobes beneath the volcano may indicate the presence of multiple magnetized bodies or distinct volcanic units contributing to the overall magnetic response. Together, these observations suggest that tectonic and magmatic processes have jointly influenced the present-day magnetic fabric of the volcanic complex.

The upward continuation results provide additional evidence regarding the scale and continuity of the magnetic sources. The persistence of the principal anomaly following 1 km upward continuation indicates that a substantial component of the magnetic field is associated with sources extending beyond the shallowest levels of the subsurface. This behavior may reflect the presence of a consolidated volcanic conduit, intrusive

complexes, or accumulations of mafic volcanic material that retain strong magnetic properties at depth. The continuity of the anomaly from shallow to deeper levels is therefore consistent with a structurally coherent magmatic system beneath the Amado Nervo Volcano.

Magnetic contrasts within the volcanic complex also provide insight into possible alteration–related processes. The subdued magnetic response observed within the central volcanic sector contrasts with the higher–intensity magnetic lobes surrounding the edifice. This pattern may reflect differences in magnetic mineral preservation, whereby hydrothermal alteration and prolonged fluid–rock interaction reduce bulk remanent magnetization in the central sector, while peripheral volcanic units retain a larger proportion of their original magnetic mineral assemblages. Similar magnetic signatures have been documented in other volcanic–hydrothermal systems where alteration is concentrated near eruptive centers and structurally controlled fluid pathways (e.g., Okubo *et al.*, 1985; Fuentes–Arreazola *et al.*, 2023).

The analytical signal results provide additional evidence. The contrast between the relatively high analytical signal amplitudes observed at the Amado Nervo Volcano and the lower values associated with the San Pedro Domes suggests significant differences in the magnetic characteristics of their subsurface sources. The lower amplitudes at the San Pedro Domes may reflect diffuse magnetic boundaries resulting from hydrothermal alteration, structural disruption, or enhanced fracturing. The spatial relationship between analytical signal boundaries and regional fault systems further suggests that tectonic structures may have influenced both the emplacement of magnetic sources and their subsequent modification. Likewise, the analytical signal responses observed at Cerro Las Tetillas and Cerro Lobos indicate that significant magnetic sources occur throughout the volcanic field. At the same time, the alignment of Cerro Lobos with the Milpillás–Cerro Grande fault highlights the potential role of regional structures in controlling volcanic emplacement.

Taken together, the magnetic anomalies, analytical signal patterns, and upward–continuation results support a model in which the interaction between magmatic and tectonic processes controls the Amado Nervo volcanic system. Regional fault systems appear to influence the geometry and distribution of magnetic sources, while alteration–related processes may locally modify magnetic properties. The persistence of major magnetic anomalies at depth, combined with evidence for magnetic attenuation in specific sectors, suggests that the volcanic complex is underlain by a long–lived magmatic framework that continues to influence its structural evolution.

The combined gravity and magnetic signatures support the interpretation of a structurally controlled volcanic framework characterized by pronounced density and magnetic contrasts.

Gravity lows, magnetic attenuation, and variations in the analytical signal delineate sectors that have experienced varying degrees of structural disruption and physical–property modification, whereas persistent magnetic highs define the broader magmatic architecture of the volcanic complex. Although the individual geophysical datasets are inherently non–unique, their spatial convergence provides a coherent framework for evaluating the hydrothermal architecture and thermal evolution of the Amado Nervo volcanic system.

5.2. Hydrothermal architecture and thermal constraints

The integrated geophysical interpretation provides the basis for evaluating the thermal structure and hydrothermal architecture of the Amado Nervo volcanic system. In particular, the spectral analysis of magnetic sources offers additional constraints on the depth distribution of thermally significant crustal domains. The estimated lower magnetic boundary ($Z_B = 7.50$ km) may approximate the Curie point depth, implying that ferromagnetic minerals lose their magnetization at relatively shallow crustal levels. Assuming this interpretation is valid, the derived geothermal gradient (70.79 °C·km^{−1}) and conductive heat flow (~ 176 mW·m^{−2}) indicate a thermal regime that is elevated relative to typical continental crust and comparable to values reported for geothermal regions within the Trans–Mexican Volcanic Belt (150 – 200 mW·m^{−2}; Prol–Ledesma and Morán–Zenteno, 2019). Furthermore, the close agreement between the spectral geothermal gradient (70.79 °C·km^{−1}) and the independent borehole–based estimate of 77.92 °C·km^{−1} reported by Espinoza–Ojeda *et al.* (2023) lends support to the reliability of the spectral approach and reinforces the interpretation of a thermally active volcanic–hydrothermal system beneath the Amado Nervo complex.

The depth distribution of magnetic sources suggests a vertically organized volcanic system extending from shallow hydrothermal domains to deeper magmatic levels. The upper magnetic domain, above approximately 1.2 km depth, may have been modified by hydrothermal alteration, oxidation, fracturing, or by volcanoclastic deposits characterized by reduced magnetic susceptibility. In contrast, the deeper magnetic sources may be associated with intrusive rocks of the Jalisco Block described by Reyes–Orozco *et al.* (2019), providing a plausible explanation for the observed magnetic layering. Although no direct geophysical evidence confirms the existence of an active magma chamber, the persistence of magnetic sources at depth, together with the elevated geothermal gradient, is consistent with the presence of a residual intrusive body that supplies heat to the hydrothermal system. Similar interpretations have been proposed for other volcanic geothermal systems where deep intrusive complexes provide long–lived thermal input despite

the absence of active silicic magma reservoirs (Henley and Ellis, 1983; Arnórsson, 1995).

The conceptual model further suggests that the transition in geophysical properties identified near sea level may correspond to an important lithological boundary separating the volcanic edifice from deeper intrusive and basement rocks. If this interface represents a significant permeability contrast, it could contribute to the development of a cap–rock system that restricts vertical fluid migration and promotes the accumulation of geothermal fluids at depth. Beneath this level, the fractured crystalline basement and associated intrusive bodies may provide the structural framework for geothermal fluid storage and circulation.

Based on the estimated geothermal gradient, temperatures of approximately 150 °C are expected at depths of 1.5–1.7 km (0–200 m.b.s.l.), near the inferred base of the cap–rock sequence. These temperatures fall within a range commonly considered suitable for geothermal development. In the proposed hydrothermal framework, meteoric recharge may occur through fractured volcanic units and fault–controlled permeable zones, allowing groundwater to penetrate toward deeper thermally anomalous sectors. Subsequent mixing between ascending thermal fluids (Heat–mass flux envelope) and cooler meteoric waters may promote cooling and vapor condensation, potentially explaining the absence of fumarolic manifestations and the occurrence of intermediate–temperature springs characteristic of secondary–condensation hydrothermal systems, identified in the vicinity of Amado Nervo Volcano (Tello–Hinojosa, 1994).

Structural controls appear to play a fundamental role in organizing the hydrothermal system. The regional Borbollón, Las Cebollas, and Milpillas–Cerro Grande fault systems define major zones of crustal weakness that likely influenced volcanic emplacement and may continue to govern fluid circulation. The coincidence of gravity lows, magnetic attenuation, and analytical signal minima with sectors affected by faulting suggests that fracturing and structural permeability are concentrated within specific portions of the volcanic complex. In this framework, meteoric recharge may occur through fault–controlled pathways and fractured volcanic units, whereas heat transfer and hydrothermal circulation are focused within structurally favorable sectors of the basement and volcanic succession.

The spatial relationship between geothermal wells and geophysical anomalies provides additional support for this interpretation. Geothermal production wells at San Pedro Domes are located along the margins of the principal gravity low, whereas geothermal–gradient wells are preferentially associated with domains characterized by subdued analytical signal amplitudes and locally elevated geothermal gradients. Although the specific causes of these geophysical signatures are not unique, their recurrent association with geothermal manifestations suggests that

density reduction, magnetic attenuation, and structural complexity collectively represent useful indicators of hydrothermal activity. Similar relationships have been documented in other volcanic geothermal systems where alteration, fracturing, and fluid circulation modify the physical properties of the host rocks (e.g., Okubo *et al.*, 1985; Arnórsson, 1995; Pandarinath *et al.*, 2019).

The overall architecture of the proposed hydrothermal system shows notable similarities to the conceptual model for the San Pedro Domes geothermal field (Reyes-Orozco *et al.*, 2019), in which hydrothermal alteration, elevated temperatures, and basement–controlled fluid circulation have also been documented. In both cases, the volcanic sequence overlies basement rocks of the Jalisco Block, and geothermal activity appears to be strongly influenced by fault–controlled permeability and lithological contrasts. This correspondence suggests that both systems may share common tectonic and thermal controls within the San Pedro–Ceboruco Graben and may form part of a broader geothermal province associated with regional extensional tectonics and late Quaternary volcanism.

Although additional drilling, petrophysical characterization, and subsurface imaging are required to validate the proposed system geometry, the integrated geophysical and geothermal evidence supports the presence of a structurally controlled volcanic–hydrothermal system characterized by elevated heat flow, shallow thermal anomalies, pervasive faulting, and extensive hydrothermal alteration. Together, these characteristics provide a coherent framework for evaluating the geothermal significance of the Amado Nervo volcanic complex and establish the basis for the geothermal potential assessment presented in the following section.

5.3. Geothermal implications, uncertainties, and limitations

The estimated geothermal potential of the Amado Nervo volcanic system should be interpreted in the context of both supporting geophysical evidence and the uncertainties associated with the adopted resource–assessment methodology. The proposed reservoir geometry is based on integrating gravity anomalies, magnetic attenuation, analytical signal patterns, geothermal–gradient estimates, and the conceptual hydrothermal model developed in this study. Collectively, these observations suggest a structurally controlled hydrothermal system characterized by enhanced fracturing, hydrothermal alteration, and elevated thermal gradients. The inferred reservoir beneath the volcanic sequence and within the fractured crystalline basement is therefore consistent with the available geophysical constraints, although its geometry and continuity remain unverified by deep drilling.

The comparison with the San Pedro Domes geothermal field provides additional evidence supporting the plausibility of the

proposed model. Similarities in structural setting, estimated reservoir temperatures, depth to basement, and evidence of hydrothermal alteration may indicate that both systems share common tectonic and thermal controls within the San Pedro–Ceboruco Graben. Nevertheless, these similarities should not be interpreted as proof of equivalent reservoir properties, as significant differences in permeability, fluid chemistry, reservoir volume, and thermal recharge may exist between the two systems.

The Heat in Place assessment provides a first-order estimate of the geothermal resource and suggests that the Amado Nervo volcanic system may possess a resource capable of supporting small- to medium-scale geothermal development. However, the uncertainty associated with the estimate remains substantial, as reflected by the wide range of simulated outcomes. This uncertainty arises from the limited availability of direct subsurface information and from the assumptions required to define key model parameters. Several of these parameters were adopted from published geothermal studies, theoretical ranges reported in the literature, or analogies with nearby geothermal systems such as San Pedro Domes. Consequently, the resulting geothermal potential should be regarded as a preliminary estimate rather than a proven resource.

Additional uncertainty stems from assumptions regarding reservoir thickness, areal extent, porosity, recovery factor, thermal properties, and reservoir temperature distribution. Moreover, the Curie–point–depth–derived geothermal gradient represents an average crustal thermal condition and may not fully capture local thermal heterogeneity associated with fault-controlled fluid circulation. Likewise, the conceptual reservoir geometry remains dependent on indirect geophysical evidence and has not yet been validated through exploratory drilling or detailed reservoir characterization.

Despite these limitations, the convergence of multiple independent datasets—including gravity anomalies, magnetic anomalies, analytical signal attenuation, Curie point depth estimates, geothermal gradients, and regional geological and hydrogeological information—provides a coherent framework supporting the geothermal prospectivity of the Amado Nervo volcanic system. The estimated resource magnitude places the system within the range reported for several medium-scale geothermal projects in Mexico, including San Pedro Domes (35.5 mW), Las Tres Vírgenes (10 MW), and Cerritos Colorados (5–15 MW). Although substantially smaller than major geothermal fields such as Cerro Prieto (500 MW) or Los Azufres (240 MW), the resource estimate suggests that Amado Nervo may represent a viable target for future geothermal exploration. Further progress will require exploratory drilling, detailed geochemical studies, reservoir modeling, and economic evaluation to reduce uncertainty and determine the technical feasibility of geothermal development.

6. Conclusions

This study integrated geological information, gravity and magnetic data, geothermal gradient measurements, and spectral magnetic analysis to develop a conceptual hydrothermal model for the Amado Nervo volcanic system. The combined geophysical evidence reveals a structurally controlled volcanic framework characterized by significant density and magnetization contrasts, suggesting fractured and hydrothermally altered domains associated with the regional fault network.

Gravity and magnetic anomalies delineate zones of reduced density and attenuated magnetization beneath the Amado Nervo Volcano, Cerro Las Tetillas, and the San Pedro Domes area. These geophysical signatures are spatially associated with major regional structures, including the Milpillas–Cerro Grande, Borbollón, and Las Cebollas fault systems, highlighting the importance of tectonic controls on volcanic evolution, structural permeability, and hydrothermal fluid circulation.

Spectral analysis of the reduced-to-the-pole magnetic anomaly yielded a Curie point depth of approximately 7.5 km, from which a geothermal gradient of $\sim 70.8\text{ }^{\circ}\text{C}\cdot\text{km}^{-1}$ and a conductive heat flow of $\sim 176\text{ mW}\cdot\text{m}^{-2}$ were estimated. These values are consistent with previously reported geothermal gradients in the region and support the existence of an elevated thermal regime beneath the Amado Nervo volcanic complex. The resulting conceptual model is compatible with a vertically organized volcanic–hydrothermal system in which a residual intrusive heat source may sustain hydrothermal activity at depth.

Integration of available geophysical, geological, and geothermal information suggests the presence of a potential geothermal reservoir beneath the inferred cap sequence rock and within the fractured crystalline basement rocks. Estimated temperatures of approximately $150\text{ }^{\circ}\text{C}$ at depths of 1.5–1.7 km indicate conditions that may be favorable for geothermal development and provide a preliminary target for future exploration drilling.

Application of the volumetric Heat in Place method coupled with 9,999 Monte Carlo simulations yielded a preliminary geothermal potential estimate of $19.75 \pm 14.41\text{ MW}$, with simulated values ranging from approximately 5 to 34 MW. Although this estimate is subject to uncertainty associated with reservoir geometry, thermal parameters, recovery factors, and the limited availability of direct subsurface information, it places the Amado Nervo volcanic system within the range of medium-scale geothermal prospects reported in Mexico.

Overall, the results support the geothermal prospectivity of the Amado Nervo volcanic complex and justify continued exploration efforts, including exploratory drilling, detailed geochemical investigations, hydrogeological characterization, and reservoir modeling. While the proposed conceptual model

remains constrained primarily by indirect geophysical evidence, the integrated methodology applied here provides a robust framework for geothermal assessment. It may be transferable to other volcanic regions of the Jalisco Block and the Trans–Mexican Volcanic Belt.

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