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Local and Duration Magnitude Scales for Northern Algeria based on Local Earthquake Data (2010-2022)

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Abstract

The study of earthquake magnitudes plays a crucial role in understanding seismic events, their impacts, and the tectonic processes that shape our planet. Earthquake magnitudes, as numerical measures of an event's size, provide valuable information about the energy released during an earthquake, helping to assess the event's potential for damage and guiding seismic hazard assessments. Over time, several magnitude scales have been developed to address different types of seismic waves and varying distances from the earthquake source. These include the Local Magnitude (M_L), Surface Wave Magnitude (M_s), Body Wave Magnitude (m_b), and Moment Magnitude (M_w), each with its advantages, limitations, and suitable applications. In northern Algeria, a region prone to seismic activity due to its complex geodynamics and seismotectonic setting, accurate magnitude determination is essential. The Local Magnitude (M_L) and Duration Magnitude (M_d) scales are of particular importance for characterizing earthquakes within the region. The M_L scale, developed by Charles Richter in the 1930s, remains one of the most widely used scales, particularly for small to moderate local earthquakes. Meanwhile, the M_d scale, which measures the duration of seismic waves, provides an important supplement to M_L , especially for larger and deeper earthquakes. Understanding these scales and their applications in northern Algeria is vital for earthquake hazard mitigation, infrastructure planning, and public safety. This thesis aims to explore and refine the calculation and use of magnitude scales, specifically focusing on the Local Magnitude (M_L) and Duration Magnitude (M_d) scales, in the context of northern Algeria. By developing a comprehensive understanding of these scales, this study contributes to the ongoing efforts to enhance seismic monitoring and improve earthquake magnitude estimation techniques in the region.

Keywords : Earthquake, seismic waves, attenuation, local magnitude, duration magnitude, northern Algeria.

Résumé

L'étude des magnitudes des tremblements de terre joue un rôle crucial dans la compréhension des événements sismiques, de leurs impacts et des processus tectoniques qui façonnent notre planète. Les magnitudes des tremblements de terre, en tant que mesures numériques de la taille d'un événement, fournissent des informations précieuses sur l'énergie libérée lors d'un séisme, ce qui aide à évaluer le potentiel de dommages de l'événement et à orienter les évaluations du risque sismique. Au fil du temps, plusieurs échelles de magnitude ont été développées pour traiter différents types d'ondes sismiques et des distances variées par rapport à la source du séisme. Parmi celles-ci, on trouve l'échelle de magnitude locale (ML), l'échelle de magnitude des ondes de surface (Ms), l'échelle de magnitude des ondes de volume (mb) et l'échelle de magnitude du moment (Mw), chacune avec ses avantages, ses limites et ses applications spécifiques.

En Algérie du Nord, une région sujette à une activité sismique importante en raison de son cadre géodynamique complexe et de ses caractéristiques sismotectoniques, la détermination précise des magnitudes est essentielle. L'échelle de magnitude locale (ML) et l'échelle de magnitude de durée (Md) revêtent une importance particulière pour caractériser les séismes dans cette région. L'échelle ML, développée par Charles Richter dans les années 1930, reste l'une des échelles les plus utilisées, en particulier pour les séismes locaux de faible à moyenne magnitude. L'échelle Md, qui mesure la durée des ondes sismiques, constitue un complément important à ML, notamment pour les séismes plus grands et plus profonds. La compréhension de ces échelles et de leurs applications en Algérie du Nord est cruciale pour l'évaluation du Aléa sismique, la planification des infrastructures et la sécurité publique.

Cette thèse vise à explorer et à affiner le calcul et l'utilisation des échelles de magnitude, en se concentrant particulièrement sur les échelles de magnitude locale (ML) et de durée (Md), dans le contexte de l'Algérie du Nord. En développant une compréhension approfondie de ces échelles,

cette étude contribue aux efforts continus pour améliorer la surveillance sismique et affiner les techniques d'estimation des magnitudes des tremblements de terre dans la région

Mots-clés : Séisme, ondes sismiques, atténuation, magnitude locale, magnitude de durée, Nord de l'Algérie.

الملخص :

يعد دراسة مقاييس الزلازل من العناصر الأساسية في فهم الأحداث الزلزالية وآثارها والعمليات التكوينية التي تشكل كوكبنا. تعتبر مقاييس الزلازل، بوصفها قياسات رقمية لحجم الحدث الزلزالي، مصادر هامة للمعلومات حول الطاقة التي يتم تحريرها خلال الزلزال، مما يساعد في تقييم حجم الأضرار المحتملة للحدث الزلزالي وتوجيه تقديرات المخاطر الزلزالية. على مر الزمن، تم تطوير عدة مقاييس لقياس الزلازل تتعامل مع أنواع مختلفة من الموجات الزلزالية والمسافات المتباينة من مصدر الزلزال. من بين هذه المقاييس نجد مقياس (mb) ، مقياس الحجم للموجات الجسدية (Ms) ، مقياس الحجم لموجات السطح (ML) الحجم المحلي ، وكل منها له مزاياه وقيوده وتطبيقاته الخاصة. في الجزائر الشمالية، التي تعد (Mw) مقياس الحجم للحظة منطقة نشطة زلزالياً بسبب الإعداد الجيوديناميكي المعقد والظروف السيزموتكتونية، فإن تحديد المقاييس من المقاييس (Md) و مقياس الحجم للمدة (ML) الدقيقة للزلازل يعد أمراً أساسياً. يعد مقياس الحجم المحلي ، الذي طوره تشارلز ريختر في ML الهامة التي تستخدم لتوصيف الزلازل في هذه المنطقة. حيث أن مقياس ثلاثينات القرن الماضي، لا يزال من أكثر المقاييس استخداماً، خاصة للزلازل المحلية التي تتراوح في قوتها ، ML ، الذي يقيس مدة الموجات الزلزالية، فيشكل مكملاً مهماً لـ Md بين المتوسطة والصغيرة. أما مقياس لا سيما بالنسبة للزلازل الكبيرة والعميقة. إن فهم هذه المقاييس وتطبيقاتها في الجزائر الشمالية أمر بالغ الأهمية لتقييم المخاطر الزلزالية، وتخطيط البنية التحتية، وضمان سلامة الجمهور. تهدف هذه الأطروحة إلى استكشاف وتحسين حساب واستخدام مقاييس الزلازل، مع التركيز بشكل خاص على مقياس الحجم المحلي ، في سياق الجزائر الشمالية. من خلال تطوير فهم عميق لهذه المقاييس، (Md) و مقياس الحجم للمدة (ML) تساهم هذه الدراسة في الجهود المستمرة لتحسين مراقبة الزلازل وتحسين تقنيات تقدير مقاييس الزلازل في المنطقة

الكلمات المفتاحية: زلزال، الموجات الزلزالية، التوهين، الحجم المحلي، حجم المدة، شمال الجزائر

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General Introduction:

Seismic hazard assessment is essential for mitigating earthquake risk, especially in tectonically active regions such as northern Algeria. The region lies at the convergence of the African and Eurasian plates, creating a highly dynamic tectonic environment characterized by compressive forces (DeMets et al., 1994; Nocquet, 2012; Bougrine et al., 2019). These forces result in active fault systems, mountainous terrain, and frequent seismic activity.

The Tellian Atlas, a significant geological feature, is the primary zone of tectonic activity and hosts the majority of seismic events in northern Algeria (Meghraoui, 1988). Tectonic Earthquakes in this region are the direct result of ongoing plate interactions, which highlight the importance of understanding the region's geodynamic framework. Broader geodynamic processes influence seismicity, and geological structures such as faults and fractures are key in governing earthquake generation and propagation. Critical considerations include the origins of seismicity in northern Algeria, faulting mechanisms, and the clustering of seismic activity along this tectonically active strip. These insights are essential for refining seismic hazard assessments, and guiding urban planning and disaster preparedness strategies in the region.

Seismic waves are vibrations generated by earthquakes that travel through the Earth's interior and along its surface. These waves are captured by specialized instruments called seismographs, which convert ground motion into a zig-zag trace that illustrates the amplitude of the Earth's oscillations at a specific location. Modern seismographs are highly sensitive and can amplify even the smallest ground motions, enabling them to detect important earthquakes occurring anywhere on the globe. By analyzing the data from multiple seismograph stations, seismologists can accurately determine the time, location, depth, and magnitude of an earthquake, as well as gain insights into the earthquake's source mechanisms and the Earth's internal structure.

In northern Algeria, the study of seismic events has advanced significantly over recent decades

(e.g., [Bezzeghoud et al., 1994](#); [Bounif et al., 2004](#); [Semmane et al., 2015](#); [Yelles-Chaouche et al., 2022](#); [Abacha et al., 2023a](#); [Abacha et al., 2023b](#); [Roubeche et al., 2024](#); [Boulahia and Semmane, 2024](#)), driven by technological improvements and a growing focus on seismic hazard mitigation. The transition from analog to digital seismic recording systems has revolutionized earthquake monitoring, providing unprecedented precision and data quality. Networks like the Algerian Permanent Seismological Network (ADSN [Yelles-Chaouche et al., 2013](#)) play a pivotal role in these advancements, offering high-quality seismic data that enhances the accuracy and reliability of observations. The deployment of advanced broadband and short-period seismic sensors, coupled with improved data transmission technologies and rigorous quality control procedures, ensures that seismic data is both precise and of high-quality. This has enabled comprehensive investigations into earthquake source mechanics (e.g., [Semmane et al., 2017](#); [Tikhamarine et al., 2024](#)), seismic wave attenuation ([Benkaci et al., 2022](#)), and local seismic tomography (e.g., [Chiarabba et al., 1997](#); [Abacha et al., 2014](#); [Bellalem et al., 2015](#); [Kherroubi et al., 2017](#)), among other topics. These studies are essential for understanding the tectonic processes shaping northern Algeria and assessing seismic hazards. They play a vital role in enhancing risk management and preparedness across the region. Spanning a broad spectrum of scientific investigations, these efforts include analyzing earthquake records (e.g., [Tikhamarine et al., 2024](#)), modeling ground motion (e.g., [Laouami, 2019](#); [Boulahia and Semmane, 2024](#)), and estimating the probability of future seismic events (e.g., [Talbi et al., 2013](#); [Rahmani et al., 2023](#)). A key parameter in these efforts is the ***Magnitude***, which plays a central role in understanding the impacts of earthquakes. The magnitude serves as a critical measure of the energy released during an earthquake and is essential for accurate seismic hazard assessments.

Before the advent of modern seismology, earthquakes were described primarily based on their visible impacts and the experiences of those who felt them—a concept now referred to as intensity. These early descriptions were qualitative, focusing on observable factors such as the intensity of ground shaking, damage to buildings, changes in the landscape, and effects on people and the environment. Observers relied on eyewitness accounts, damage reports, and the geographic extent of shaking to gauge an earthquake’s severity. Over time, intensity scales were developed to systematically categorize these effects. For example, the Modified Mercalli Intensity Scale ([Wood and Neumann, 1931](#)) classified earthquakes based on observed damage and human perception of shaking, providing a relative measure of an earthquake’s impact at specific locations. While inten-

sity measurements effectively conveyed the extent of destruction and shaking, they only reflected localized effects rather than the total energy released by the earthquake. This limitation highlighted the need for a more scientific and quantitative approach to earthquake measurement, leading to the development of magnitude scales, which measure the energy released at the source of the earthquake.

The magnitude of an earthquake is a numerical value that relates to the amplitude of seismic motion. However, estimating magnitude is complex and can vary depending on the methods and datasets used. For instance, while the "Richter scale" (commonly referred to as the *local magnitude* or M_L) is frequently mentioned in the media, it is not always the most appropriate scale for large or distant earthquakes. In fact, reported magnitudes may change as more data becomes available, highlighting the complexity of earthquake magnitude estimation. Numerous magnitude scales have been developed over the years. The first magnitude scale, introduced by [Richter, 1935](#), measures the maximum recorded signal amplitude for observations within 600 km of the earthquake. While M_L was initially designed for earthquakes in Southern California, recorded on the Wood-Anderson torsion seismometer ([Anderson and Wood, 1925](#)), it continues to be widely used in local and regional seismic catalogs (eg., [Hutton and Boore, 1987](#); [Alsaker et al., 1991](#); [Bindi et al., 2007](#); [Saunders et al., 2013](#)). However, M_L has limitations: it underestimates magnitudes for distant earthquakes (>600 km), deep earthquakes, and large events, due to its inability to fully account for attenuation, surface waves, and shaking duration.

To address these limitations, other magnitude scales were developed to complement M_L : The duration magnitude M_d which is based on the duration of seismic shaking, provides an indirect measure of the energy released during an earthquake. It is particularly useful for smaller earthquakes, where amplitude measurements may be less reliable. The surface-wave magnitude (M_s), introduced by [Gutenberg, 1945a](#), is based on the maximum amplitude of surface waves with a period of about 20 seconds. This scale was developed to better characterize large, shallow earthquakes at regional and teleseismic distances. Later, [Gutenberg, 1945b](#) introduced the body-wave magnitude (m_b), which is calculated from the amplitude and period of seismic body waves, typically P-waves with a period of about 1 second. While these scales were initially adjusted to align with M_L for moderate earthquakes, they all share a common limitation: saturation at large earthquake magnitudes.

Another class of magnitude scales is based on the seismic moment. The moment magnitude (M_w), introduced by [Kanamori, 1979](#), has become the most physically meaningful magnitude scale. Derived from the seismic moment (M_0), M_w does not saturate at large magnitudes and is directly related to the geometric and physical properties of the earthquake fault, such as rupture area, slip, and shear modulus. Despite its disadvantages, M_L remains crucial for seismic monitoring, especially in the creation of seismic catalogs that help identify areas of high seismicity. However, linking narrowband magnitudes like M_L to broadband scales like M_w often introduces significant biases, emphasizing the need for careful calibration.

In northern Algeria, the Earthquake Monitoring Department at the Center for Research in Astronomy, Astrophysics, and Geophysics (CRAAG), along with studies by Algerian researchers (e.g., [Semmane et al., 2012](#); [Yelles-Chaouche et al., 2014](#)), has utilized an unpublished relationship inspired by [Lee et al., 1972](#) to compute the duration magnitude (M_d). This magnitude scale was recently adopted for compiling the new Algerian seismic catalog, covering the period from 2006—when the Algerian Digital Seismic Network (ADSN) was established—to 2020, as documented by [Yelles-Chaouche et al., 2022](#). However, Algeria has not yet developed specific local and duration magnitude scales tailored to its region. Accurate earthquake magnitude estimation is central to seismic hazard assessment, serving as a primary measure of an earthquake's size and impact. In Northern Algeria, developing a local (M_L) and duration (M_d) magnitude scales tailored to the region's specific geological conditions is essential for improving hazard assessments. Precise magnitude estimates allow for better characterization of seismic events and support the development of more reliable seismic hazard models.

The aim of this thesis is to estimate these magnitude scales by performing amplitude and duration inversions of earthquakes recorded by the ADSN in and around northern Algeria. The study focuses on seismic events that occurred between January 1, 2010, when the network's current geometry was finalized, and June 1, 2022.

This thesis is structured into five chapters:

Chapter 1 "Geological setting in northern Algeria". This chapter examines the complex geodynamic processes shaping northern Algeria and the broader Western Mediterranean region. It investigates the geological structures that contribute to the area's seismicity and aims to answer key questions, including the origins of Algerian seismic activity and the reasons behind its concentra-

tion in the northern strip of the country.

Chapter 2 "Earthquake magnitude and intensity scales". This chapter explores earthquake measurements, including intensity and magnitude, and provides an overview of the various magnitude scales that have been developed over time. Particular emphasis is placed on the methodology used in estimating local magnitude (M_L) and duration magnitude (M_d), highlighting the calculation techniques and considerations specific to these scales.

Chapter 3 "Data and Methods". This chapter provides a comprehensive overview of the data acquisition and analysis methods employed throughout the thesis. The focus is on the Algerian Permanent Seismological Network, which serves as the foundation for continuous, real-time seismic data collection essential for the research. The chapter details the data preparation workflow, beginning with the installation of seismic stations and the retrieval of raw data. Key preprocessing steps, such as time corrections, deconvolution, and filtering, are described, leading to the organization of the data into a coherent, structured database for further analysis.

Chapter 4 "A Local magnitude scale (M_L) for northern Algeria". This chapter focuses on the development of a local magnitude scale (M_L) for northern Algeria. The analysis is based on 17,377 seismic records from 1,901 high-quality earthquakes. Events were carefully selected based on criteria such as signal-to-noise ratio (SNR), time intervals between events, and maximum depth. The instrument response was deconvolved, and the resulting waveforms were convolved with a simulated Wood-Anderson (WA) instrument response to accurately determine maximum displacement amplitudes. The study also examines amplitude decay as a function of distance and evaluates the suitability of different attenuation models (single, bilinear, or trilinear). This led to the derivation of a distance correction formula and the establishment of a M_L - M_w conversion relationship using data from the Algerian network. The research improves magnitude estimation accuracy, contributing to seismic risk reduction and the protection of vulnerable communities in the region.

Chapter 5 "A Duration magnitude scale (M_d) for northern Algeria". This chapter will focus into the estimation of the duration magnitude (M_d) for a series of seismotectonic events in northern Algeria. The study is based on a dataset of 1,901 earthquakes recorded at 72 stations within the Algerian Digital Seismic Network (ADSN) between January 2010 and June 2022. It explores the development of a duration magnitude scale specific to the region, building on the local

magnitude scale and improving the understanding of seismic events in northern Algeria.

Finally, refining magnitude scales and improving our understanding of seismic hazards are critical for more accurate predictions of earthquake impacts. These advancements contribute to the development of resilient infrastructure, effective emergency response strategies, and targeted risk reduction initiatives. For northern Algeria, tailoring magnitude scales to account for local seismicity ensures that seismic hazard assessments are both better assessed and region-specific. Furthermore, the methodologies and insights gained from these efforts have broader applications for other seismically active regions, particularly those with similar tectonic settings. By integrating region-specific and global seismological approaches, we can support both local and international efforts in seismic hazard mitigation.

Chapter 1

Geological setting in northern Algeria

Understanding the geodynamic and seismotectonic context of Algeria is crucial for assessing seismic hazards and enhancing seismic monitoring capabilities in this tectonically active region. Algeria, located at the boundary between the African and Eurasian plates, is subject to significant tectonic forces that have shaped its geomorphological features, including mountainous landscapes, Neogene basins, and active fault systems (Meghraoui, 1988; Abacha, 2015). These tectonic interactions continue to generate seismic events, particularly concentrated along the northern region, predominantly within the Tellian Atlas strip (Yelles-Chaouche et al., 2022). This chapter delves into the intricate Geological setting in Northern Algeria and the broader Western Mediterranean, investigating the geological structures that underlie the region's seismicity. It seeks to answer key questions such as the origins of Algerian seismicity and the reasons behind its concentration in the northern strip. The geological and seismotectonic framework of Algeria directly impacts the accuracy of local and duration magnitude scales. Variations in crustal composition, fault geometry, and seismic wave propagation characteristics across the region necessitate region-specific calibrations for reliable magnitude estimation (Hutton and Boore, 1987). Northern Algeria's complex subsurface structures, such as the Tellian Atlas's fold-and-thrust systems, strike-slip faults networks, and the extensive Neogene sedimentary basins, introduce heterogeneities in seismic wave attenuation, which must be accounted for in magnitude calculations. Without these corrections, systematic biases may arise, leading to inaccurate assessments of earthquake size and associated hazards. Furthermore, the high seismicity of the northern regions presents an ideal natural laboratory to refine magnitude scales tailored to tectonic environments where earthquake activity is

spatially heterogeneous. For instance, the proximity of seismic sources to densely populated areas amplifies the importance of precise magnitude determination to improve early warning systems and seismic risk assessments. The influence of site effects, including local geology and soil conditions, further underscores the need to integrate seismotectonic and geodynamic considerations into magnitude estimations (Saunders et al., 2013). By exploring the interplay between geological structures and seismic activity, this chapter provides a foundation for understanding how regional characteristics influence the parameters used in seismic monitoring. These insights contribute to more robust magnitude scales, ultimately supporting the broader goals of seismic hazard mitigation and preparedness in Algeria.

1.1 General geology

Northern Algeria consists of several geological and structural units, forming part of the Maghreb Alpine chain that stretches from Gibraltar to Calabria (Durand-Delga, 1969) (Figure 1.2b). Within Algeria, this chain can be divided into distinct regions: to the north, the Tellian chain or the Tell, which includes the Kabylia (Lesser and Greater Kabylia that formed the internal zone), flyschs, and Tellian nappes, and to the south, the intra-continental Atlas system (Saharan Atlas, Aures Massif, and High Plateaus) (Figure 1.1).

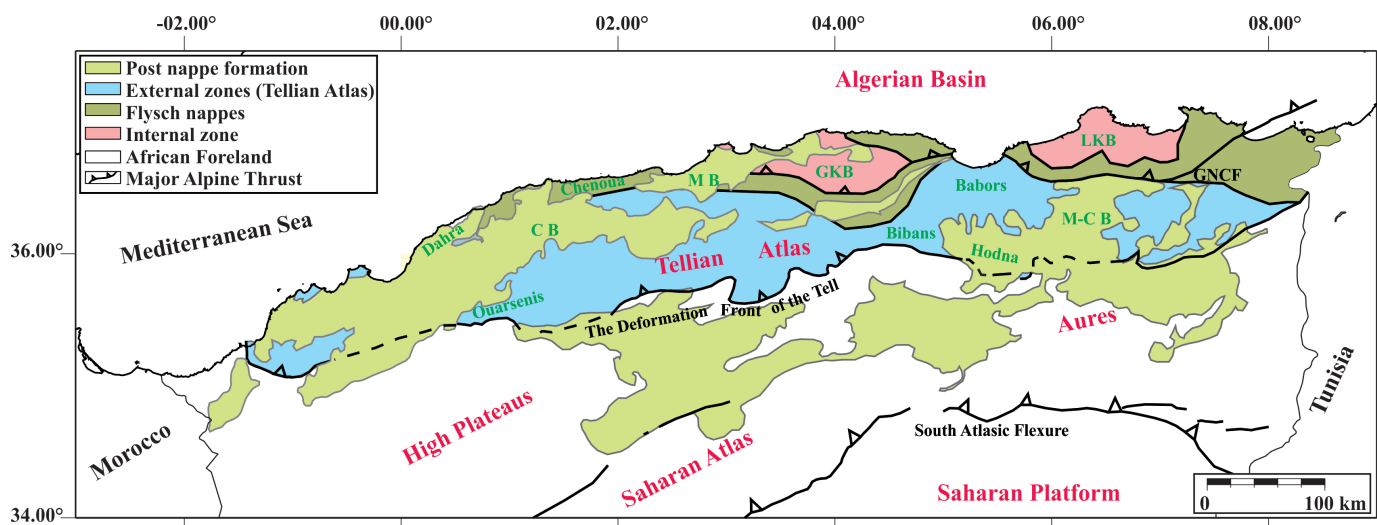


Figure 1.1: Map of the major geological domains of northern Algeria, adapted from Yelles-Chaouche et al., 2022.

The Kabylia, or internal zones of the northern Tell, primarily encompass Greater and Lesser Kabylia Blocks (GKB and LKB, Figure 1.4). These regions are composed of a Paleozoic crystalline metamorphic basement, originating from the fragmentation of the AlKaPeCa. Additionally, the Kabylia Ridge (Dorsal), a Meso-Cenozoic limestone chain, is divided into three main units: internal, median, and external (Durand-Delga, 1969). **The Flysch nappes**, composed of conglomerates and other detrital materials, occupy various positions: internally, they are back-thrust onto the internal zones (northern Kabylia flyschs); in a relatively external position, they are located along the southern edge of the Kabyle Dorsal (southern Kabylia flyschs); or they can be completely external, thrust up to a hundred kilometers southward onto the external Tell (Figure 1.4). **The Tellian nappes**, or external zones of the southern Tell, consist of imbricated allochthonous nappes (e.g. Babors; Bibans;...) (Figure 1.1), primarily composed of marls ranging from the Triassic to the Neogene age. These nappes originated from the African paleomargin and have been thrust southward, sometimes over a hundred kilometers. They are divided into three major units (Villa, 1980): (1) the ultra-Tellian units; (2) the Tellian s.s. units; and (3) the peni-Tellian units. South of the Tell, the intracontinental Atlas, separated from the Saharan platform to the south by the South Atlasic Flexure, is composed of autochthonous units. It is divided into several regions: the Saharan Atlas in the center and the Aures Massif in the east (Figure 1.1). The High Plateaus, located in western Algeria between the Tell (South of the deformation front of the Tell) and the Saharan Atlas mountains, consist of a Paleozoic basement with minimal Meso-Cenozoic sedimentary cover and limited deformation. South of the Atlas lies the Saharan platform, composed of a Precambrian basement covered by relatively undeformed Paleozoic and Mesozoic sediments. Finally, in the area of the Algerian submarine margin (Algerian basin in Figure 1.2b), which is heavily influenced by deposits from the Messinian Salinity Crisis (Lofi et al., 2011).

1.1.1 Geodynamics and Seismotectonics

Algeria is located in North Africa and lies at the boundary between the African and Eurasian plates in the Western Mediterranean region. The NW-SE convergence of these two plates is related to the counter-clockwise rotation (red arrows in Figure 1.2a) of Africa relative to Eurasia, with a Euler pole located off the coast of Morocco (red point in Figure 1.2a) (DeMets et al., 1994). Convergence directions in the western Mediterranean region vary between N300°-350°, with a variable

rate of 2.2 to 8.3 mm/yr (Nocquet, 2012; Bougrine et al., 2019) (Figure 1.2b). This convergence has resulted in the formation of relief features such as the mountain ranges of the Tellian Atlas (e.g., Hodna, Bibans, Babors, Dahra, Chenoua, Blida Atlas) and the Saharan Atlas, with the High Plateaus lying in between (Figure 1.2c). Seismic activity is primarily occurring along the Algerian margin from north to south, including the Tellian Atlas and related Neogene basins such as the Cheliff, Mitidja, and Mila-Constantine basins (CB, MB, and M-CB respectively, in Figure 1.2c) (e.g., Meghraoui, 1988; Boudiaf, 1996; Abacha, 2015). Interestingly, seismicity is less pronounced within the Greater and Lesser Kabylia Blocks (GKB and LKB in Figure 1.2c and Kabylia in Figure 1.2b), resulting from the breakup of the AlKaPeCa blocks (for Alboran, Kabylia, Peloritan and Calabria). Yelles-Chaouche et al., 2021 explain this phenomenon to the rigidity of the blocks due to their Paleozoic geological formation. Seismic activity is less affected by present active deformation further south, in the High Plateaus or the Saharan Atlas (Figure 1.2c and 1.2d). Crustal thickness is estimated to be approximately 20–30 km along the Algerian continental margin and the Tellian Atlas, 30–36 km in the High Plateaus region, and even thicker in the Saharan Atlas, at around 36–44 km (Melouk et al., 2022). This study utilized data from 15 Algerian broadband seismic stations (Figure 1.3). Additionally, a complementary recent survey conducted by Guettouche et al., 2024, using eight three-component short-period stations, estimated the mean Moho depth in the Guelma pull-apart basin and its surroundings—a region not covered by the first study—to be between 22 and 28 km (Blue box in Figure 1.3). The horizontal stress axes (SHmax) (Yelles-Chaouche et al., 2022, and references therein) and the velocity field orientation (Bougrine et al., 2019, Figure 1.2d) in central and western Algeria exhibit greater uniformity and homogeneity compared to the eastern region. The narrow coastal strip (50–100 km wide) in this region, where active thrust faulting and seismicity are concentrated, indicates deformation within a compressive stress regime (domain 1). This area is particularly susceptible to high-magnitude earthquakes, such as the 1980 El Asnam earthquake (Ms 7.3) (e.g., Ouyed et al., 1981) and the 2003 Boumerdes earthquake (Mw 6.9) (e.g., Ayadi et al., 2003) (Figure 1.2c and 1.2d). In contrast, the eastern region, which accounts for approximately two-thirds of Algeria's seismicity, is the most seismically active part of the country (Abacha, 2015) (Figure 1.2d). This region experiences widely distributed deformation characterized by low to moderate-magnitude seismic sequences, commonly referred to as seismic swarms. In a recent study (Rahmani et al., under review in *Natural Hazards*), 32 swarms were recorded in

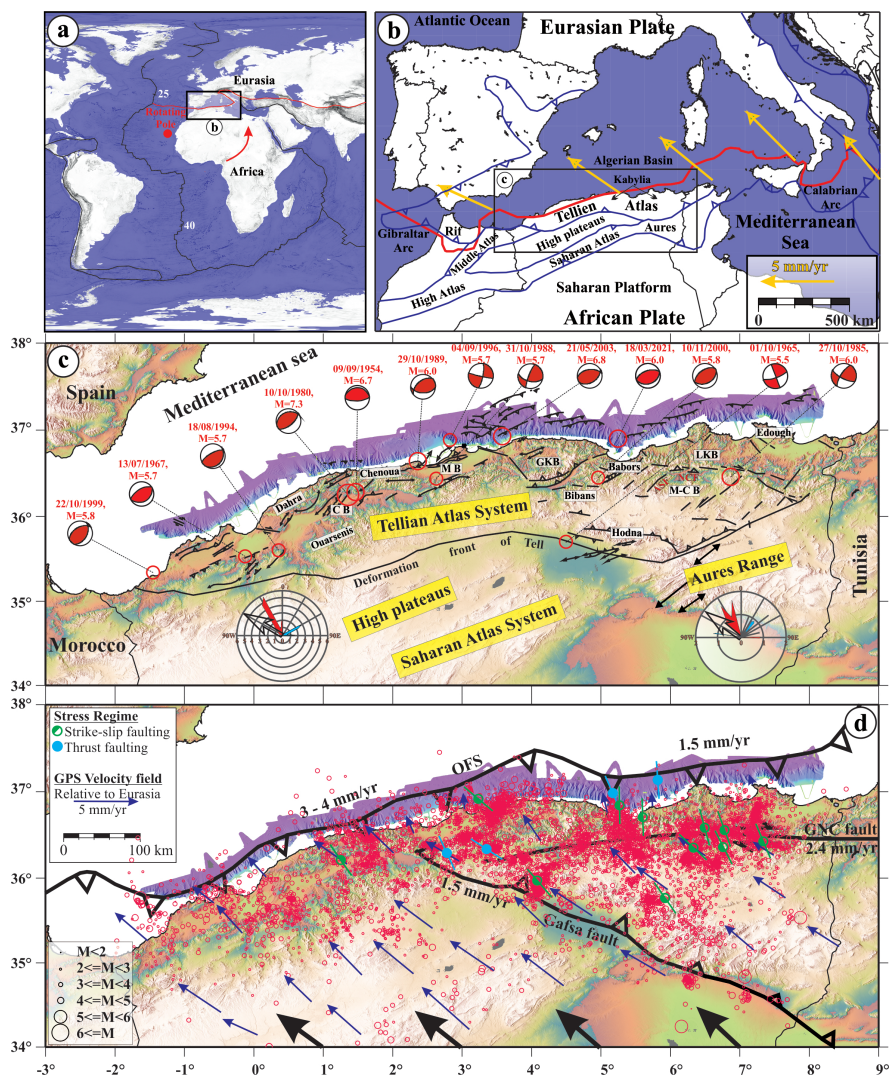


Figure 1.2: **(a)** Current configuration of tectonic plates (e.g., Africa and Eurasia) modified from [Nocquet, 2012](#). The numbers on the ridges indicate the relative velocities between the plates. The red line marks the boundary between the Eurasian and African plates, outlined in a broad manner; further specifics can be found in **(b)** (enclosed within the black box). **(b)** Predicted velocities along the Eurasia–Africa Plate boundary (orange arrows) are simplified from [Nocquet, 2012](#). The main geological units of Algeria are also shown. Maghrebe chain = Kabylia + Rif + Tellian Atlas. Intra-continental Atlas = Aures + Saharan Atlas [Durand-Delga, 1969](#). The black box outlines northern Algeria. The red line represents the Eurasia–Africa Plate boundary (simplified from [Bird, 2003](#)). **(c)** Seismotectonic map of northern Algeria, showing the main active faults observed in the Tellian Atlas (from [Abacha, 2015](#) and references therein), as well as the large earthquakes ($M \geq 5.5$) and their focal mechanisms. The Neogene seismogenic basins (CB: Cheliff basin, MB: Mitidja basin, and M-CB: Mila-Constantine basin) and the main mountain ranges of the Tellian Atlas are labeled. The GKB and LKB correspond to the Greater and Lesser Kabylia Blocks, respectively. The rose diagrams show, for each domain 1 and 2, the main Pressure (P) (in red) and Tension T (in blue) axes directions were taken and modified from [Serpelloni et al., 2007](#). **(d)** Recent northern Algeria seismicity between 2006 and 2020 (CRAAG catalog). SHmax obtained from the inversion of the focal mechanisms of the main recent seismic events. The key structures and the velocities field are taken from [Bougrine et al., 2019](#).

this eastern region between 2007 and 2020. These seismic sequences are associated with strike-slip faulting. The faults involved include left-lateral NE–SW faults, right-lateral NW–SE faults, and right-lateral WNW–ESE faults, the most significant of which are the Ghardimaou–North Constantine (GNC) Fault (or M'cid Aïcha-Debbagh Fault) (Abacha and Yelles-Chaouche, 2019; Boulahia et al., 2021; Bendjama et al., 2021; Abacha et al., 2023b) and the Beni-Illmane fault networks (Yelles-Chaouche et al., 2014 Beldjoudi et al., 2016; Abacha et al., 2023a; Tikhamarine et al., 2024). Comparison of the mean orientation of the SHmax and velocity vectors with the direction

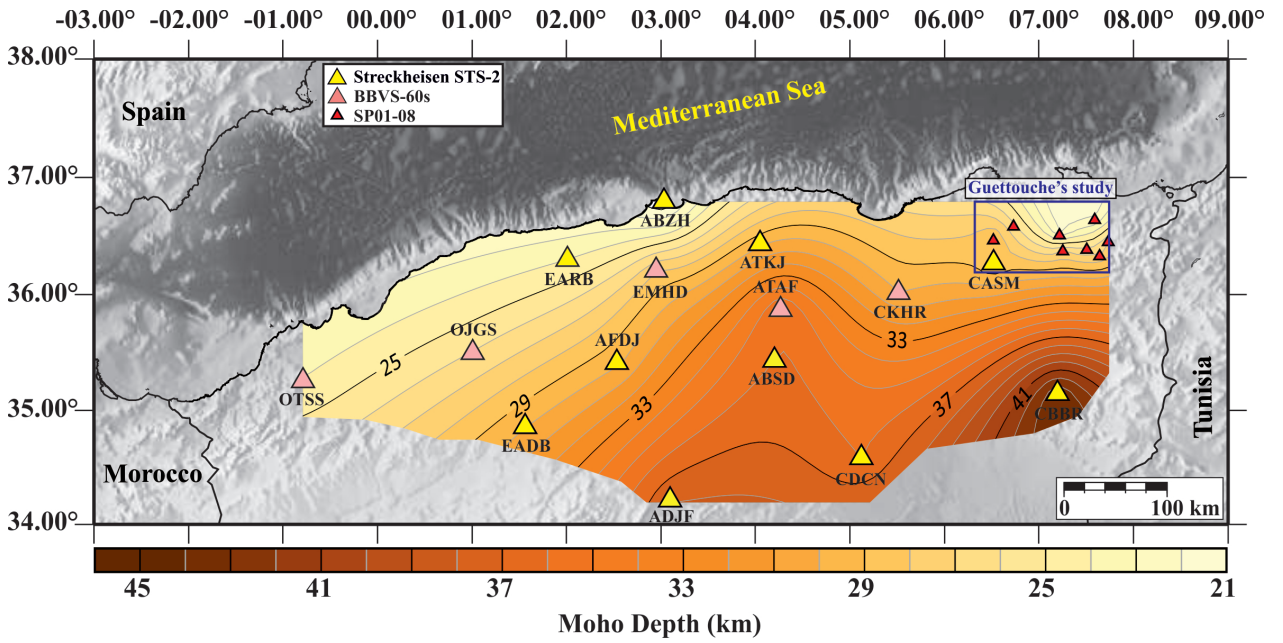


Figure 1.3: The Moho contour map (modified from Guettouche et al., 2024) includes the results derived from Melouk et al., 2022, along with the eastern extension provided by Guettouche et al., 2024.

of relative movement of the African and Eurasian plates (Figure 1.2d), confirms the homogeneous nature of central and western Algeria. This observation aligns with findings by Soumaya et al., 2018 (Figure 1.4). However, the eastern region (domain 2) shows a reorientation of the stress and velocity field, particularly in the far northeast, around the LKB, influenced by the role of the GNC fault (Boulahia et al., 2021; Bendjama et al., 2021; Yelles-Chaouche et al., 2021; Yelles-Chaouche et al., 2023). Bougrine et al. [2019] revealed that oblique convergence in the eastern region is partitioned into three main regions: (1) the offshore thrust fault system with a rate of 1.5 mm/year, (2) the GNC fault with a rate of 2.4 mm/year, and (3) the Gafsa fault with a rate of 1.5 mm/year. This heterogeneous zone can increase the seismic attenuation. Conversely, in the western region,

no such partitioning occurs, and the convergence is accommodated in shortening faults and thrust folds with a rate between 3 to 4 mm/year (Figure 1.2d).

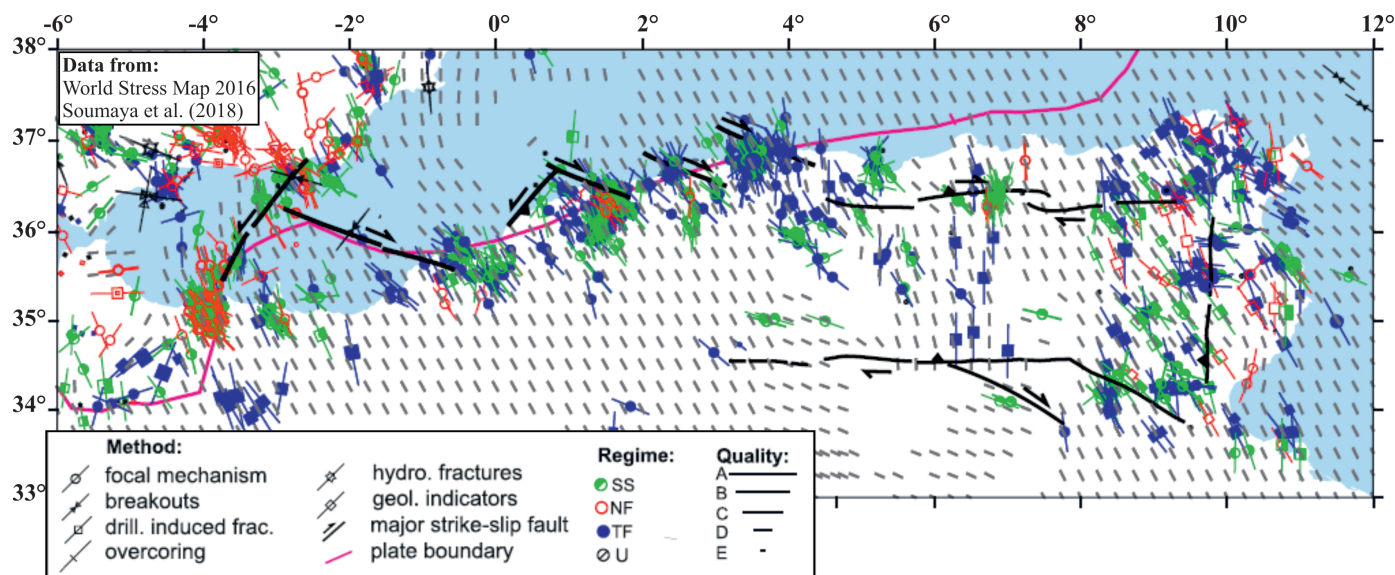


Figure 1.4: Stress map and smoothed stress pattern modified from Soumaya et al., 2018 of the Maghreb region. Different lines show the orientation of the SHmax.

1.1.2 Geomorphological setting

From a geomorphological perspective (Figure 1.1), northern Algeria, features diverse landforms comprising both basins and mountain ranges. These structures reflect the region's complex geological and tectonic evolution.

1- Basins

- **Constantine and Mila Intermontane Basins:** These basins are located amidst mountainous terrains, characterized by high altitudes, steep slopes, and narrow, deeply incised valleys.
- **Annaba Neogene and Quaternary Basin:** A sedimentary basin composed of Neogene and Quaternary deposits, it is situated along the northeastern coastal area of Algeria, reflecting active tectonics and sedimentation processes.
- **Guelma Basin:** This pull-apart basin is bounded by E-W dextral strike-slip faults. Its tectonic origin and structural features make it an important area for understanding fault-controlled sedimentation.

- **Soummam Basin:** This narrow E-W to SW-NE-oriented depression exhibits features of tectonic subsidence, with significant fluvial sedimentation along its axis.
- **Hodna Neogene Basin:** Characterized by chotts (salt lakes), this basin is a prominent feature of the central region, reflecting both evaporitic processes and tectonic activity.
- **Chott Melrhir Basin:** Structured during the Tertiary period, this basin features a substantial Cretaceous fill, with sedimentary deposits reaching thicknesses of up to 5000 m.
- **Mitidja Basin:** Located south of Algiers, this geological depression is oriented ENE-WSW. It spans approximately 120 km in length and 20 km in width, situated at the foothills of the Atlas Mountains. The Mitidja Basin is primarily filled with Quaternary and Neogene sediments.
- **Chelif Basin:** Found in northwestern Algeria, this basin covers an area of approximately 45,000 km^2 . It contains marine and continental deposits ranging from the Upper Miocene to the Pliocene, highlighting its significance in regional sedimentary and tectonic studies.

2- Mountain Ranges

- **Babors Chain:** This fold belt, composed predominantly of limestone, trends WSW-ENE and is distinguished by its high mountains and deeply incised valleys.
- **Bibans Chain:** Predominantly E-W oriented, this range is composed of Tellian nappes, representing a tectonically deformed structure.
- **Djurdjura Chain (Kabylia Ridge or Dorsal):** A prominent limestone range trending E-W and NO-SE. Its high elevations and karstic features make it a significant geomorphological and tectonic landmark.
- **Hodna Mountains:** Characterized by reliefs oriented N110, this range forms part of the central Algerian landscape.
- **Aurès Mountains:** These mountains exhibit a predominant NE-SW orientation and represent the southern extension of the Saharan Atlas.

- **Nemencha Mountains:** Similar to the Aurès range, these mountains are oriented NE-SW and are marked by rugged terrains and tectonic deformation.
- **Ouled Naïl Mountains:** Also trending NE-SW, this range is known for its complex relief and structural features.
- **Ouarsenis Massif:** Spanning approximately 200 km from east to west, this mountainous region showcases diverse landscapes, from forested areas to steep, rocky slopes.
- **Chenoua Massif:** This Hercynian Alpine massif, located along the North African margin, is one of the few remnants of the Hercynian orogeny, making it a key site for geological and tectonic research.
- **Dahra Mountains:** Known for their rugged relief and complex geological composition, the Dahra range highlights the interplay of tectonic and erosional processes.

This geomorphological diversity, encompassing intermontane basins and tectonically uplifted mountain ranges, highlights the dynamic geological and tectonic processes that continue to shape northern Algeria. These features not only provide critical insights into the region's geodynamic evolution but also serve as key determinants in understanding its seismicity and tectonic framework. Furthermore, they play a pivotal role in influencing seismic wave propagation, including attenuation patterns, site effects, and topographic amplification, all of which are essential for accurate seismic hazard assessment, modeling, and calculation of magnitude formulas.

1.2 Main factors influencing magnitude estimation

Accurate estimation of local magnitude (M_L) and duration magnitude (M_d) necessitates a thorough understanding of geological, seismotectonic, and geomorphological features, as these factors play a crucial role in shaping seismic wave propagation, attenuation, and amplification. This section highlights the key parameters influencing M_L and M_d calculations, focusing on two primary effects: attenuation in hard rock media and amplification in soft media. These factors collectively determine the accuracy of magnitude assessments and provide essential insights into regional seismic characteristics.

1.2.1 Geological factors

- **Crustal Structure:** The thickness and composition of the Earth's crust play a critical role in seismic wave propagation. Thin crusts, as exemplified by the northern part of the coastline in Figure 1.3, tend to amplify seismic waves. In contrast, thick crusts, (Figure 1.3), generally attenuate wave amplitudes. Additionally, geological features like sedimentary basins or crystalline basement layers introduce heterogeneities that significantly affect wave propagation patterns, altering their speed, direction, and energy distribution.
- **Seismic Velocity Models:** Variations in seismic velocities significantly affect wave travel times and amplitudes:
 - **P-wave velocity (V_p):** Low velocities in sedimentary basins delay wave travel and reduce amplitudes, whereas high velocities in hard rock formations enhance wave propagation.
 - **S-wave velocity (V_s):** Differences in V_s influence the V_p/V_s ratio, which is crucial for travel time corrections and magnitude calculations.
 - **V_p/V_s Ratio:** Low values (e.g., < 1.7) indicate brittle, dry crust, while higher ratios suggest the presence of fluids or partial melts, increasing attenuation.

In Figure 1.5a, the seismic tomography of northern Algeria is presented in terms of P-wave velocity anomalies, derived from the work of [Aichaoui et al., 2022](#) using teleseismic events. In this representation, high-velocity (hard rock) regions are shown in blue, while low-velocity (soft rock) regions are depicted in red. Figure 1.5b displays the results of a local earthquake tomography study ([Abacha et al., 2025, in revision](#)), in which I am a co-author. These panels illustrate P-wave and S-wave velocity anomalies, as well as V_p/V_s ratio distributions. Elevated V_p/V_s values in this context likely indicate the presence of fluid reservoirs, providing valuable insights into subsurface geological and tectonic processes.

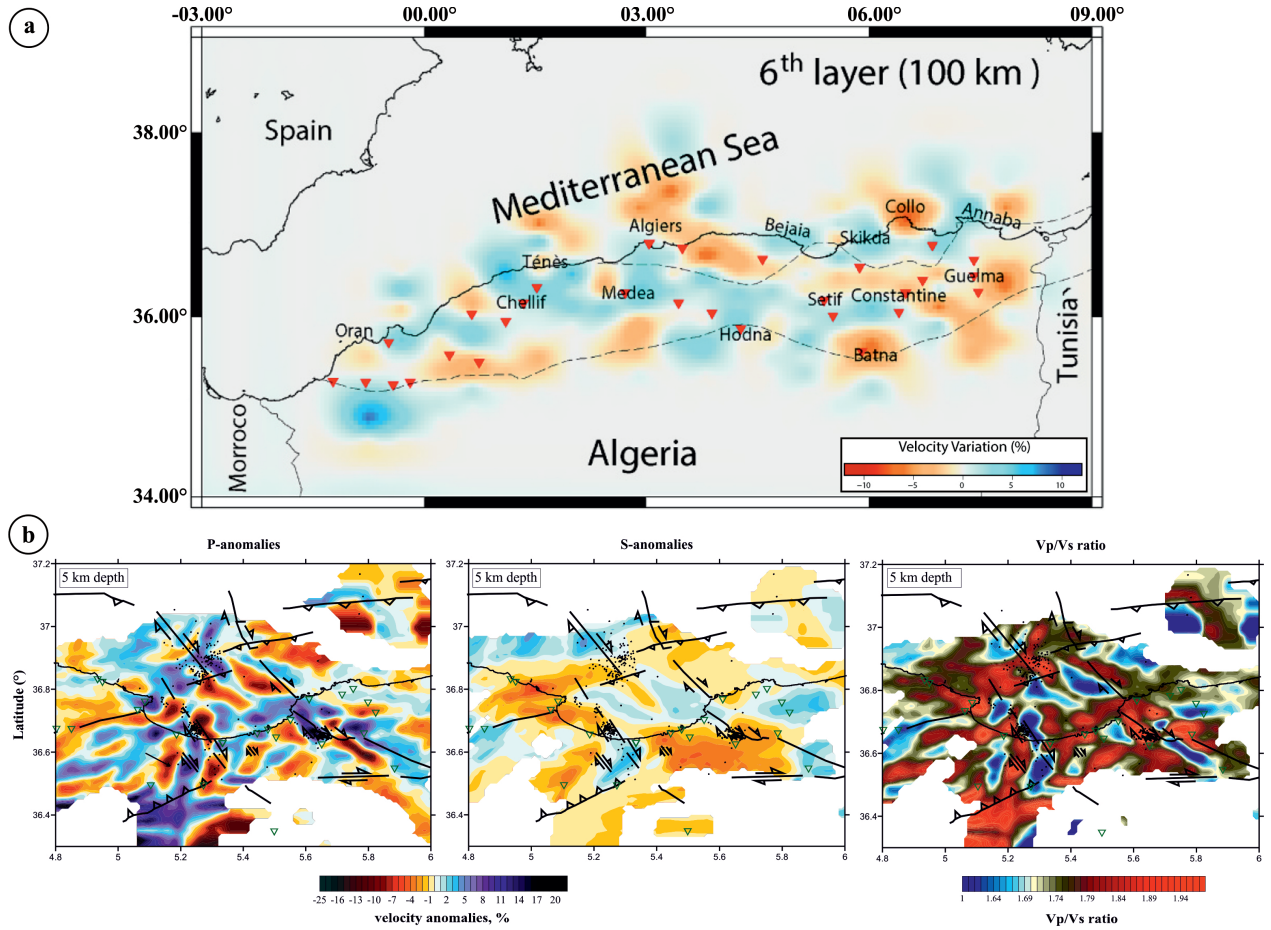


Figure 1.5: (a) Maps of P-wave velocity variations across Northern Algeria at a depth of 100 km from regional earthquake tomography (Aichaoui et al., 2022). (b) Horizontal sections showing velocity anomalies of P- and S-waves in Bejaia-Babors region, along with Vp/Vs ratios, at a depth of 5 km from local earthquake tomography (Abacha et al., 2025, in revision).

- **Attenuation Characteristics:** The attenuation of seismic waves, quantified by the quality factor Q (the inverse of the attenuation), controls the energy loss due to absorption and scattering. Regions with high Q values preserve wave amplitudes over greater distances, while areas with low Q exhibit significant amplitude reduction. Scattering caused by geological heterogeneities, such as faults and fractures, further prolongs the coda duration, impacting M_d estimates.
- **Surface Geology Near-surface conditions influence seismic wave behavior:** Soft soils amplify wave amplitudes, leading to higher M_L and longer M_d . Hard rock reduces amplification, resulting in lower magnitudes. Sedimentary basins trap seismic energy, prolonging

coda waves and affecting M_d estimation (Figure 1.6a). In Figure 1.6b, we present an example of a site effects investigation conducted in the city of Chellif (Layadi et al., 2016). This study, based on earthquake and ambient vibration data, illustrates the spatial distribution of fundamental frequencies (f_0) and their corresponding amplitudes (A_0).

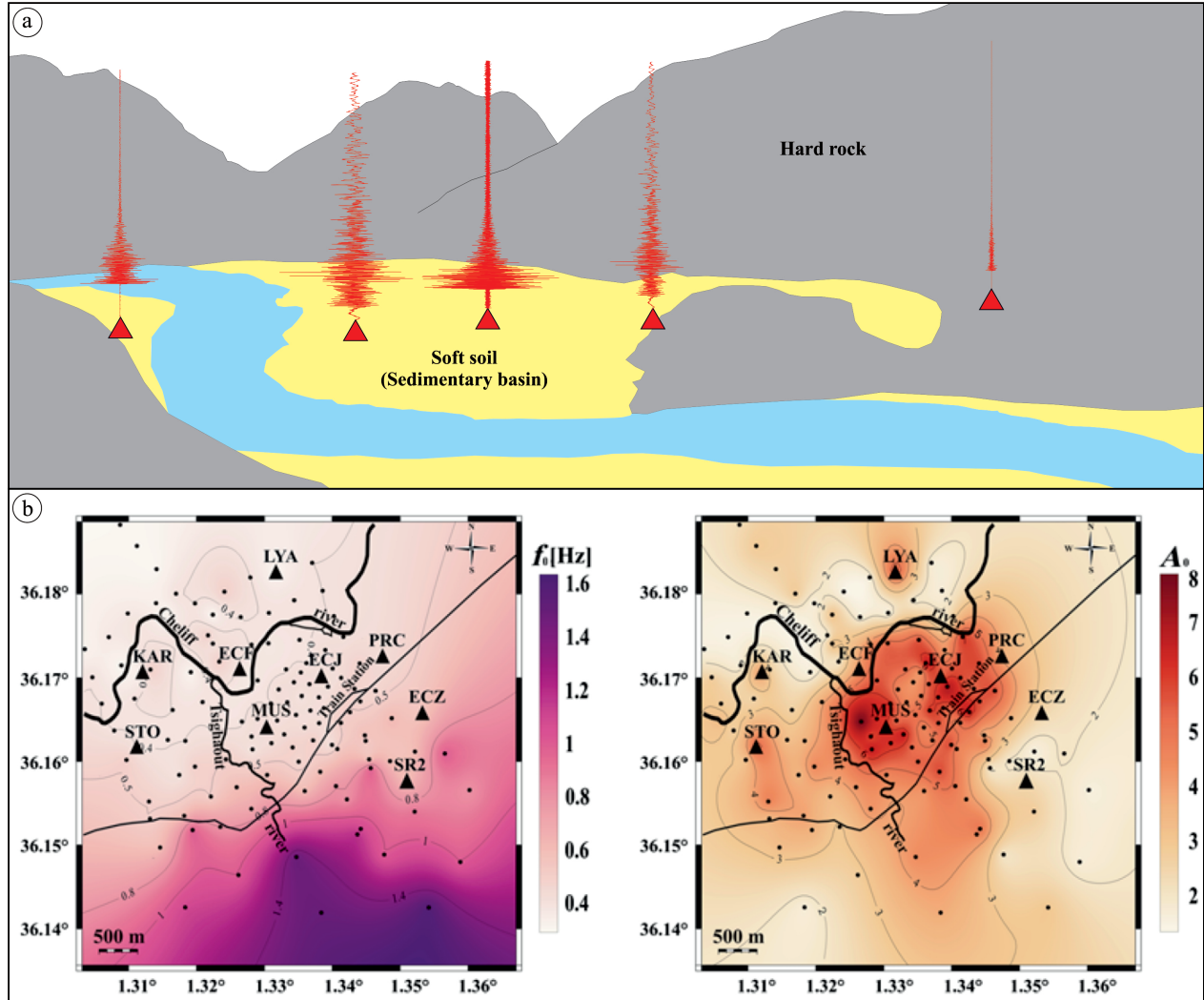


Figure 1.6: (a) Illustration of site effects in a sedimentary basin: seismogram amplitudes and durations are significantly higher in sedimentary layers compared to bedrock. (b) Contour map of fundamental frequencies of the soil (f_0) and corresponding amplitudes (A_0) in the Chellif Basin (western Algeria) obtained using earthquake and ambient vibration data, adapted from Layadi et al., 2016.

- **Topography:** Topographical variations, such as mountains or valleys, affect wave propagation. Mountainous regions can reflect and refract seismic waves, distorting their amplitudes, while flat terrains facilitate more uniform propagation with minimal distortion (Figure 1.7a).

[Layadi et al. \[2023\]](#) investigated topographic site effects at Algerian broadband seismic stations using earthquake and ambient vibration data (Figure 1.7b, top). Their analysis of directional amplification at certain stations (Figure 1.7b, bottom) revealed that amplification at hilltops or slopes tends to align perpendicularly to the major ridge axis.

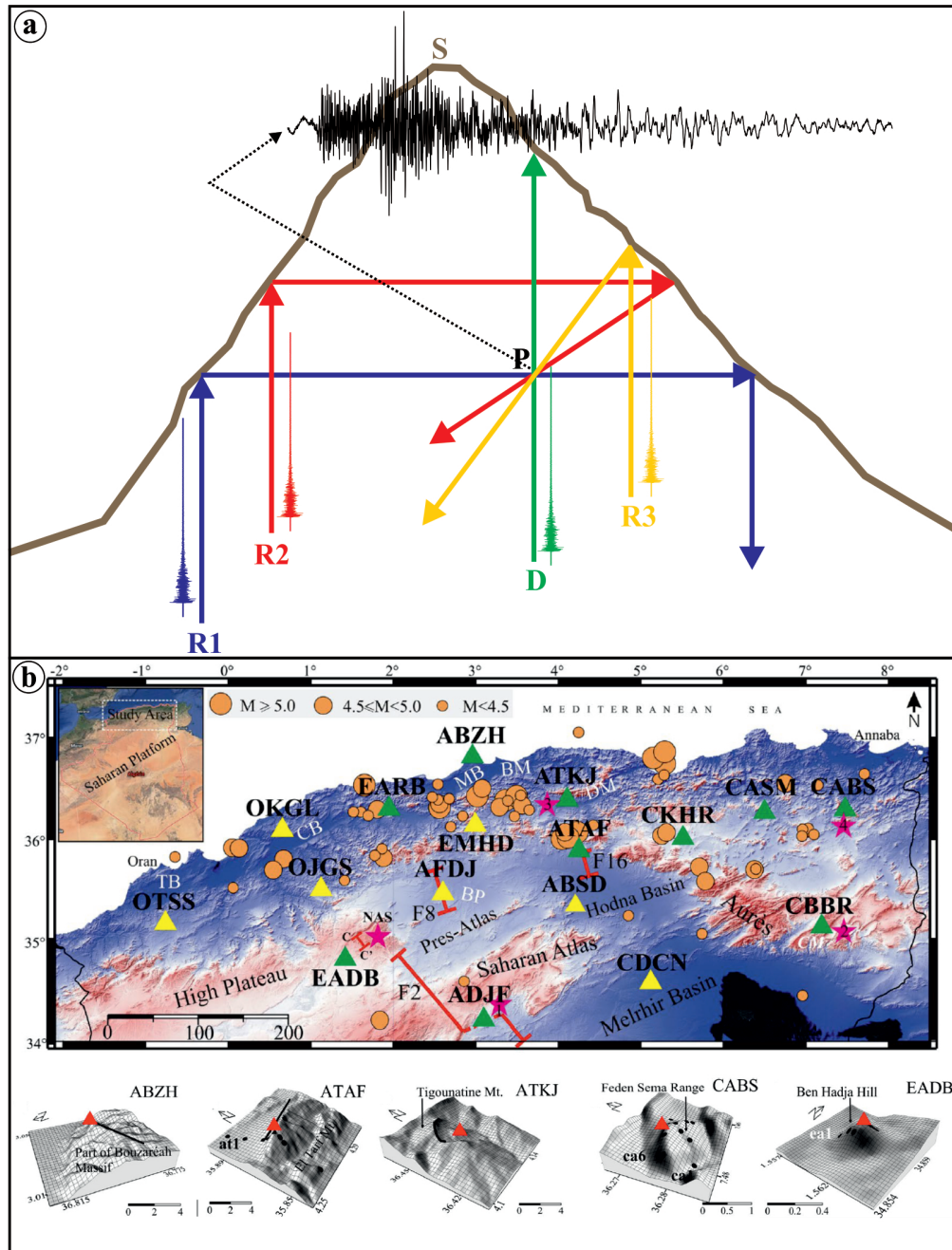


Figure 1.7: Topographic effect (the paths are drawn for convenience). (a) Reflection of a vertically propagating plane wave by a triangular hill with apex S, focusing waves at point P. (b) Topographic distribution of broadband seismological stations (red triangles) relative to the highest elevations. Solid black circles indicate ambient vibration measurement points, and black continuous lines represent the major ridge axis (Layadi et al., 2023).

- **Anisotropy:** The alignment of minerals or geological structures induces anisotropy, leading to directional variations in seismic wave velocities. This effect is particularly significant in

regions with well-developed fault systems or pronounced sedimentary layering, where azimuthal amplitude variations can influence local magnitude (M_L) estimates. On a regional scale, [Radi et al., 2023](#) identified upper mantle anisotropy beneath northern Algeria through shear wave splitting analysis (Figure 1.8). Such anisotropy may affect the magnitude estimation of both regional earthquakes and teleseismic events.

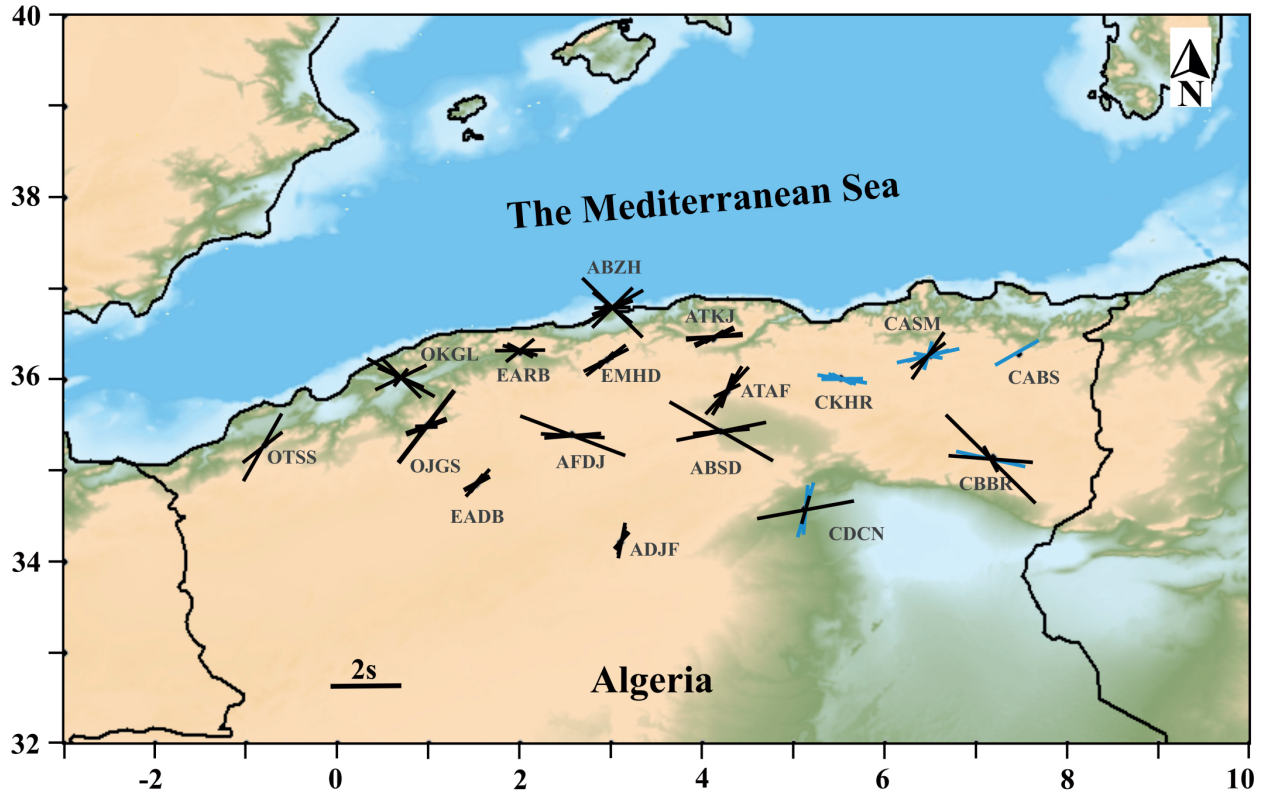


Figure 1.8: Map showing the orientation of fast SKS waves (seismic anisotropy) recorded at Algerian Broadband Seismological Stations. Blue lines represent data reported by [Radi et al. \[2015\]](#), while the black lines indicate results obtained in the study of [Radi et al., 2023](#).

1.2.2 Seismic factors

- **Fault Geometry and Activity:** As we are using data from local earthquakes, the geometry and faults activity influence seismic wave characteristics: (1) Fault Type: Strike-slip faults radiate more horizontal energy, while thrust faults generate significant vertical energy, affecting amplitude-based M_L calculations. (2) Activity Levels: Active faults with frequent ruptures increase scattering and attenuation due to fault-zone complexities. As highlighted in Section 1.1.1 and Figure 1.2, the northeastern part of Algeria exhibits swarm seismicity char-

acterized by high event productivity, short return periods, and a complex network of small fault segments, predominantly associated with strike-slip faults. In contrast, the offshore region and the northwestern Algeria part are distinguished by seismicity of greater magnitude, longer return periods, and are also primarily driven by thrust faulting. These distinct seismic characteristics must be carefully considered when estimating earthquake magnitudes.

- **Tectonic Regime:** Tectonic settings determine stress distribution and energy release patterns: (1) Compressional regimes are associated with higher stress drops, producing larger magnitudes. (2) Extensional regimes typically exhibit lower stress drops, resulting in lower M_L values. (3) Strike-slip regimes generally produce moderate stress drops, lower than those in compressional settings but higher than in extensional ones. As demonstrated in Section 1.1.1 and illustrated in Figure 1.2d, the eastern part of northern Algeria is dominated by a strike-slip regime, whereas the western part is characterized by a compressive regime. This tectonic variation is expected to significantly influence magnitude estimation due to the differing stress drops and energy release patterns associated with these regimes.
- **Seismicity Distribution:** The spatial and temporal distribution of seismic events impacts magnitude estimation: (1) Clustered Seismicity: Localized seismicity (e.g., near volcanoes) may experience greater energy dissipation due to geological complexities, reducing amplitudes and prolonging coda durations. (2) Diffuse Seismicity: Uniformly distributed seismic events yield more consistent magnitude estimates. As previously discussed, these patterns are influenced by fault geometry, tectonic regimes, and forcing mechanisms, such as fluid involvement, as illustrated below.
- **Earthquake Source Characteristics:** The depth and nature of earthquake sources affect seismic wave energy: (1) Source Depth: Shallow earthquakes (≤ 10 km) generate stronger surface waves and higher amplitudes, leading to higher M_L . Conversely, deeper events (≥ 50 km) exhibit reduced amplitudes and shorter coda durations (Lay and Wallace, 1995). In Algeria, the seismogenic layer is predominantly confined to depths between 0 and 20 km, as shown in Figure 1.9. (2) Stress Drop: High stress-drop earthquakes release more seismic energy, resulting in stronger wave amplitudes and higher M_L , which have a strong link with tectonic regimes and fluid circulation.

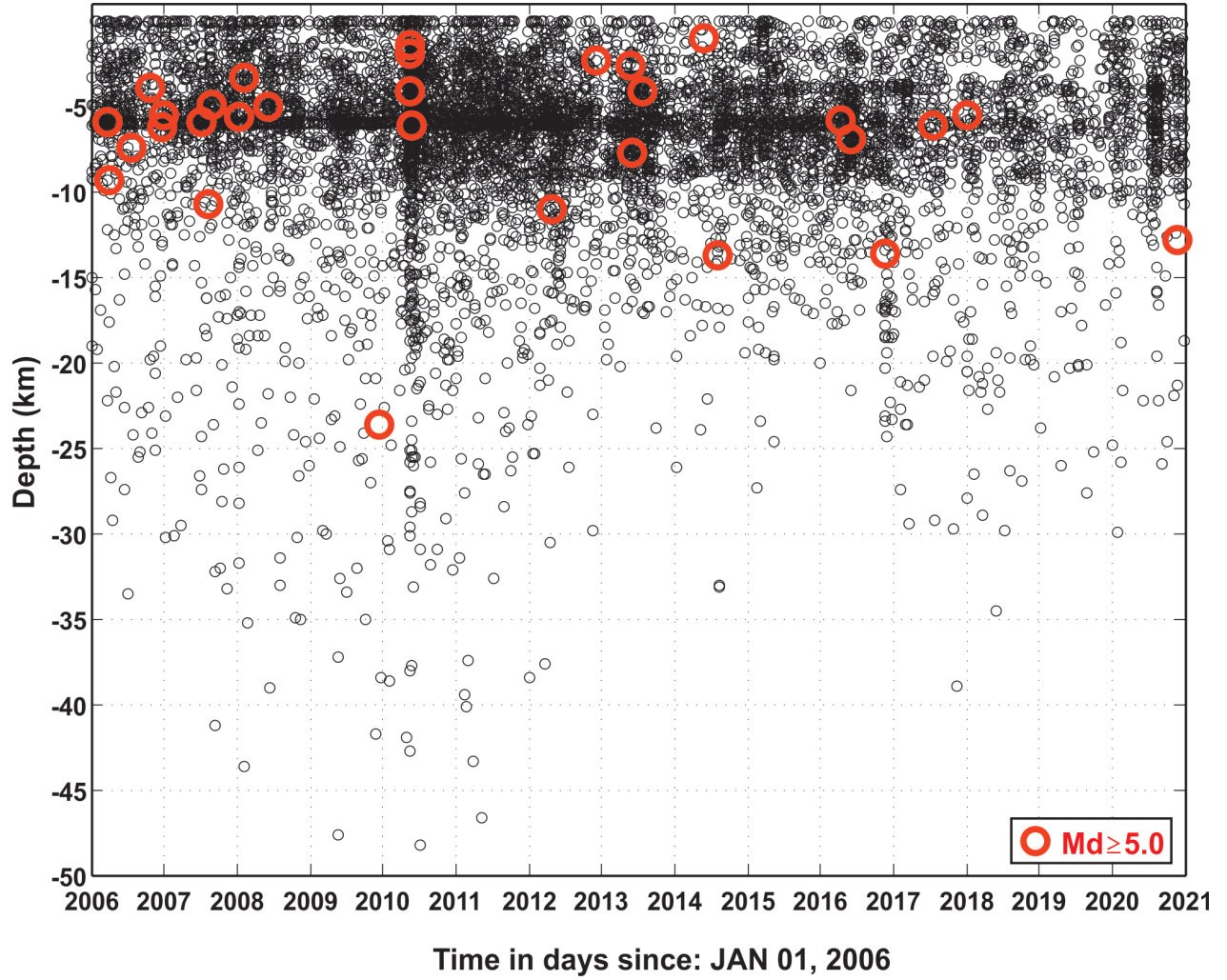


Figure 1.9: The temporal evolution of hypocenter depths in the Algerian catalog from 2006 to 2020, based on raw data, with events of magnitude 5 highlighted as circled dots (from [Yelles-Chaouche et al., 2022](#)). The location uncertainties are less than 2 km in both the horizontal (ERH) and vertical (ERZ) directions, with $RMS < 0.3$ and an azimuthal gap of less than 180° .

- **Regional Attenuation Laws:** Ground motion attenuation varies regionally, influenced by geological and tectonic factors. Empirical attenuation relationships must be calibrated for local conditions to ensure accurate M_L estimation. Areas with high attenuation often underestimate M_L for distant events.
- **Fluid Presence:** Fluids within fault zones or surrounding rocks alter seismic wave propagation: (1) High pore pressure reduces effective stress, resulting in low stress-drop events with diminished energy release ([McGarr, 2014](#)). (2) Seismic Attenuation: Fluids increase absorption and scattering, reducing amplitudes and extending coda durations. The V_p/V_s ratio is

notably sensitive to fluid saturation within rocks, serving as a key indicator of the crust's mechanical and petrological properties (Garcia et al., 2019, Ozer et al., 2018). Seismic tomography studies are highly effective tools for identifying and understanding these physical characteristics.

1.3 Conclusion

The geodynamic and seismotectonic setting of Algeria is shaped by the complex interplay between the African and Eurasian plates, giving rise to a diverse seismic landscape. Seismic activity is predominantly concentrated in the northern region, particularly along the Tellian Atlas system, where active fault systems and geological formations reflect ongoing tectonic processes. These features contribute significantly to the seismic hazards in Northern Algeria. Local magnitude (M_L) is highly sensitive to variations in wave amplitudes, which are influenced by geological heterogeneities, seismic attenuation, and site effects. Similarly, duration magnitude (M_d) depends on the length of coda waves, which are governed by scattering, attenuation, and source depth characteristics. Fluid presence, tectonic regime, fault geometry, and the spatial distribution of seismicity further modulate magnitude estimates, underscoring the need for a region-specific approach. To enhance the accuracy of M_L and M_d estimations, it is imperative to integrate region-specific seismic velocity and attenuation models that account for Algeria's unique geological, seismotectonic, and geomorphological parameters. Incorporating these factors will refine magnitude scales and significantly improve seismic hazard assessments, contributing to better disaster preparedness and mitigation strategies in the region.

Chapter 2

Earthquake magnitude and intensity scales

Understanding earthquake strength is essential for assessing potential impacts and informing engineering design to enhance resilience against seismic events. In the case of an earthquake, two key measures are typically considered: seismic intensity and magnitude, each capturing different aspects of an earthquake's impact and underlying mechanics. Seismic intensity specifically measures the strength of ground shaking at a given location. It reflects how strongly the earth's surface moves in response to seismic waves and provides a direct indication of the potential for damage to buildings, infrastructure, and landscapes. Intensity measurements are essential for evaluating localized effects of earthquakes on communities, helping to assess the need for emergency response, guide urban planning, and design structures to withstand the expected level of shaking in different regions. Notably, seismic intensity varies from one location to another based on factors like distance from the epicenter, local geological conditions, and building practices. However, while intensity is invaluable for understanding local impacts, it does not provide information about the earthquake's source size or the total energy released. For these broader aspects, seismologists turn to other measures, such as magnitude.

The concept of earthquake magnitude is a crucial advancement in seismology. It enables a quantitative comprehension of earthquake size, making it the initial parameter generally established after localizing an earthquake (determining the epicenter's geographical coordinates, hypocenter depth, and origin time). The magnitude is the most commonly used and determined parameter to characterize an earthquake. It is an important part of many scientific activities, such as the creation of earthquake catalogs, used in statistical analysis of earthquakes, and is a crucial aspect of various

scientific pursuits. These studies are fundamental to seismic hazard assessment and the calculation of parameters such as the Gutenberg-Richter b value, as shown in research (e.g., [Wiemer and Wyss, 2002](#); [Wyss et al., 2004](#); [Ghosh et al., 2008](#)).

In an ideal scenario of a perfectly elastic and homogeneous medium, seismic waves conserve their energy during propagation. Geometric spreading, where the spherical wave front expands, causes attenuation in amplitude over distance, resulting in a decrease in intensity as the wave moves through the Earth. This decrease, independent of frequency, is directly proportional to the inverse of the distance between the earthquake source and the seismic station for body waves. However, the Earth is not elastic or homogeneous; rather discontinuities, diverse rock types, and varying velocities and densities at different depths characterize it. The complex structure of the Earth's crust adds to the complexity of seismic wave propagation, resulting in amplitude decay at different rates depending on regional structural variations. Geometric spreading and scattering contribute to this decay, however, the presence of anelastic processes, referred to as anelastic attenuation, further amplifies the energy loss ([Richter, 1935](#)). The combination of elastic processes (geometric spreading and scattering) and anelastic processes (anelastic attenuation) results in the attenuation of the seismic waveform. Richter's definition recognizes the attenuation as a result of epicentral distance and hypocentral depth, illustrating the complex interplay of the Earth's varied subsurface on the reduction of seismic amplitudes.

In 1935, when Charles Richter developed his original magnitude scale, his goal was to create an empirical method to classify the relative size of small to moderate earthquakes in California. He aimed to provide a standardized measure that could distinguish the strength of local earthquakes in a straightforward, comparative way. The method involved comparing variations in maximum amplitudes recorded at a network of seismic stations, each equipped with standard short-period Wood-Anderson (WA) Torsion seismometers. Richter solved the problem of amplitude decrease with distance by incorporating an appropriate correction term for distance. Given the wide range of earthquake magnitudes and their resulting amplitudes, Richter introduced the local magnitude scale M_L . This scale is derived from seismic records obtained at epicentral distances up to 600 km and provides a comprehensive representation of earthquake magnitudes. This scale is designed to evaluate local earthquakes and includes negative magnitudes down to about -2, while the upper limit for magnitudes determined by the M_L scale is typically about seven ([Bormann and Saul,](#)

2009). Subsequently, it was discovered that magnitudes derived from short-period waves exhibit saturation. Thus, the development of complementary magnitude scales became imperative to provide magnitude estimates that were less sensitive to saturation or no saturation. A few decades ago, analog seismic recordings were dominant, characterized by a relatively limited dynamic range of about 40 dB. This limitation resulted in frequent instances where recording traces would go out of scale, especially when recording larger seismic events at local or regional distances. This made it impossible to measure the maximum amplitude. However, researchers discovered that the duration of the coda following the maximum amplitude, which is characterized by exponentially decaying amplitudes, shows an increase that correlates with both magnitude and distance. Based on this observation, formulas for the duration magnitude (M_d) were developed.

In 1954, Gutenberg [1945a] formulated the magnitude scale M_s specifically for teleseismic surface waves. This scale is based on measurements of the maximum horizontal displacement amplitudes of "true" ground motions. This maximum corresponds to the Airy phase, which represents a local minimum in the group velocity dispersion curve of Rayleigh surface waves and is attributed to the presence of a low-velocity layer in the upper mantle.

Simultaneously, in the same year, (Gutenberg, 1945b; Gutenberg and Richter, 1945) introduced a magnitude relationship for teleseismic body waves (m_b) such as P, PP, and S. This relationship is based on theoretical amplitude calculations that take into account geometric propagation and distance-dependent attenuation. These calculations are further refined by comparison with empirical observations of shallow- and deep-focus earthquakes, primarily in medium-period records. The exploration of magnitudes for significant earthquakes has consistently shown varying degrees of saturation, which has led to the proposal of a non-saturating magnitude, known as M_w (Kanamori and Stewart, 1977; Hanks and Kanamori, 1979). The moment magnitude (M_w) calculation is based on the scalar seismic moment (M_0), which quantifies the total amount of anelastic energy necessary to rupture and displace the fault associated with an earthquake. M_0 can be determined through time-domain waveform analysis and inversion or by assessing the spectral amplitude of the low-frequency level (plateau) in the P- or S-wave displacement spectrum. M_w is expected to show no saturation, as all other scales depend on amplitude or duration at fixed periods well beyond the magnitude-dependent corner period of the seismic source spectrum.

2.1 Intensity scales

Another measure of earthquake strength is seismic intensity, which describes the local severity of ground shaking based on observed damage to buildings and the experiences of people who felt the earthquake. Unlike magnitude, which quantifies the earthquake's energy at its source, intensity reflects the effects at specific locations and can vary widely across different areas. The impact of an earthquake can vary significantly from one location to another. For example, while City A may experience collapsed buildings, City B might only report broken windows, and City C might see no visible damage at all. These differences in observed effects highlight how earthquake intensity is not only influenced by the earthquake's magnitude but also by local conditions, building resilience, and personal interpretations of the damage. Because intensity assessments are based on these observed impacts, they are inherently subjective and can differ widely across regions affected by the same event. In the following part, we will focus on the most important intensity scales used in Algeria and worldwide.

Rossi-Forel Intensity Scale: The first widely accepted earthquake intensity scale was introduced in the 1880s by Italian scientist Michele Stefano De Rossi and Swiss scientist François-Alphonse Forel. The Rossi-Forel scale rates the intensity of an earthquake on a scale from I to X, reflecting the observed damage and human perception of shaking. Though it has largely been replaced by more refined scales, it is still occasionally referenced in describing earthquake effects. Notably, the 1906 San Francisco earthquake was evaluated using this scale.

Modified Mercalli (MM) Intensity Scale: Italian seismologist Giuseppe Mercalli revised the Rossi-Forel scale in 1902, expanding its descriptive power. This scale was further refined by Cancani and Sieberg in 1923, resulting in the Mercalli-Cancani-Sieberg (MCS) scale, which extended the scale to 12 degrees (I to XII) to better capture variations in earthquake effects. Later, Wood and Neumann adapted this version in the U.S., creating the Modified Mercalli (MM) scale, which became widely used for assessing earthquake intensity. A 1956 revision by Richter provided a version known as the MM scale (1956 version), which remains popular alongside the Richter scale (M_L) for assessing the severity of ground shaking.

European Macroseismic Scale (EMS): The European Macroseismic Scale (EMS-98) is a detailed and comprehensive tool for assessing earthquake intensity, enhanced by its inclusion of

guidelines, illustrations, and practical examples. Unlike traditional intensity scales such as the Modified Mercalli, which estimate building strength based on construction type alone, the EMS-98 introduces six vulnerability classes that assess structural resilience more directly. These classes account for factors such as construction type, workmanship, and condition, making EMS-98 adaptable to a variety of building types, including earthquake-resistant structures. Damage levels are categorized into five distinct grades, from "negligible to slight" to "destruction," and are illustrated for clarity.

The EMS-98 also highlights the probabilistic nature of earthquake intensity through quantitative definitions, which align with observed damage patterns across buildings of similar strength. Unlike older scales, the EMS-98 downplays natural geological effects like landslides and rockfalls—which can be influenced by non-seismic factors—by placing them in an annex. First published in 1998, the EMS-98 scale has been used effectively not only in Europe but also in assessing earthquakes worldwide.

Medvedev-Sponheuer-Karnik (MSK) Intensity Scale: Further revisions to the MCS and MM scales led to the MSK scale, developed by Soviet and European seismologists Medvedev, Sponheuer, and Karnik. Approved at a UNESCO meeting in 1964, the MSK scale aimed to offer a more detailed and objective approach. However, its complexity limited practical use. To address this, the European Seismological Commission (ESC) revised the scale in 1988, finalizing an updated version in 1992 at the XXIII ESC General Assembly in Prague. In Algeria, earth scientists used the MSK scale until recently, when the EMS-98 scale was adopted ([Harbi et al., 2007](#)) to map earthquake intensity and its correlation with local infrastructure damage, similar to the MM scale in having 12 degrees, is typically used in a more practical range from intensity III to X, as lower (I and II) and higher (XI and XII) intensities are less frequently applied in practice.

Practical Use and Limitations of Intensity Scales: While intensity scales theoretically range from I (not felt) to XII (total destruction), in practice, certain degrees are rarely used. Intensity I (not felt) and II (too weak to report) are seldom recorded, and intensities XI and XII are challenging to distinguish in practical scenarios. As a result, intensity scales effectively operate between degrees III and X, providing a range that covers the most observable earthquake effects.

2.1.1 Steps in Macroseismic Studies

When studying an earthquake's macroseismic effects (those observed over a broad area), the following steps are generally followed:

Data Acquisition: Information is collected through questionnaires, field surveys, public appeals, literature reviews, and other methods.

Data Sorting: The acquired data is organized by location and summarized to facilitate interpretation, often by grouping responses from similar areas.

Intensity Assignment: Using an intensity scale, a preliminary intensity rating is assigned for each location, and a summary table is created to reflect the intensity levels across affected areas.

A modern approach for rapidly estimating earthquake intensities is the "Did You Feel It?" platform maintained by the United States Geological Survey (USGS). This website allows users to contribute to earthquake data by answering a brief set of questions about how strongly they felt the shaking and providing their location. This crowdsourced information helps build detailed, real-time maps of perceived earthquake intensity across affected regions. For more information, visit the "Did You Feel It?" page at earthquake.usgs.gov/eqcenter/dyfi.php.

2.1.2 Isoseismal Mapping

Isoseismals are contour lines on a map that connect points experiencing the same intensity of shaking (e.g., Figure 2.1), often forming elliptical shapes that reflect the geological structures influencing the earthquake's impact. These maps can highlight regions with different intensity levels (e.g., IV or V), marking them with intensity numbers. Historically, the area of maximum intensity (the meizoseismal area) was often used to approximate the epicenter in the absence of instrumental data. However, with modern seismograph data, it's observed that the epicenter may actually lie outside the meizoseismal area. Isoseismal maps are valuable tools for geologists, who study them to understand faulting patterns, and engineers, who use them to assess the performance of construction types in affected areas. These maps also serve as critical guides for identifying weak or hazardous zones, informing safer future construction practices.

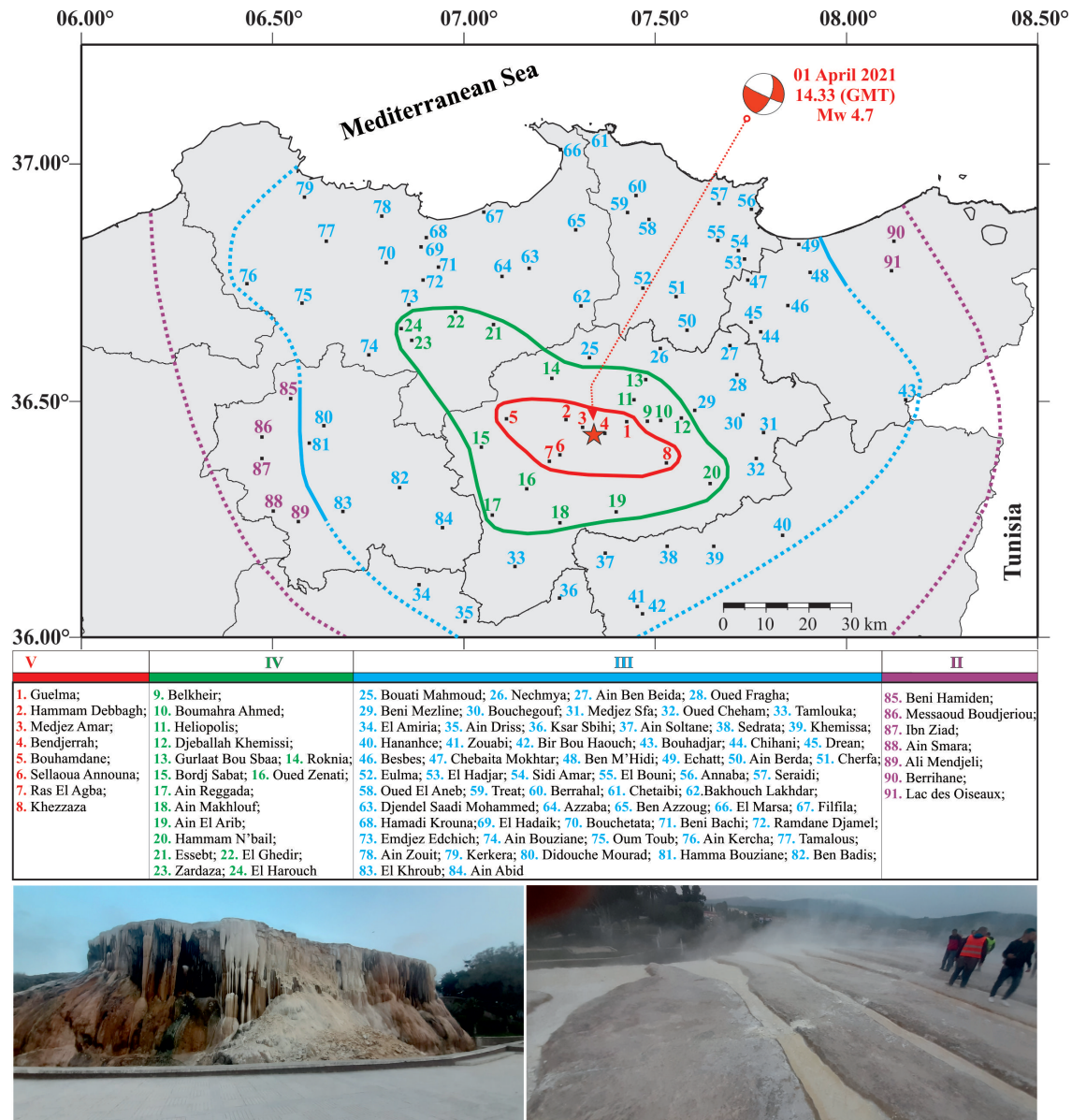


Figure 2.1: Example of an isoseismal map for the 2021 Guelma earthquake (from [Boulahia, 2022](#)).

2.2 Magnitude scales

2.2.1 Local magnitude (M_L)

Earthquake magnitude is an easy and intuitive way to compare the relative sizes of earthquakes. The commonly used scale, called local magnitude (M_L) (also known as the 'Richter Scale') was established by [Richter, 1935](#). The M_L scale is determined by a mathematical formula based on measurements taken from seismographs:

$$M_L = \log_{10}(A) - \log_{10}(A_0) \quad (2.1)$$

In this equation, A represents the maximum horizontal displacement measured in millimeters (mm) using a Wood-Anderson (WA) torsion seismometer, either as zero-to-peak or half peak-to-peak. The term $-\log_{10}(A_0)$ provides a distance correction term, taking into account the combined effects of geometric and anelastic attenuation. To date, the local magnitude scale remains the preferred choice for assessing the magnitude of local and small earthquakes, primarily because of its simplicity and speed of measurement. To adapt the M_L scale for use with instruments other than the Wood-Anderson seismometer and in regions beyond Southern California, the following M_L calculation procedure is presented:

1. **Deconvolution:** Deconvolve the instrument response of the recording sensor from the recorded waveform.
2. **Convolution:** Convolve the waveform that has been corrected using instruments from the first step with the instrument response characteristic of the Wood-Anderson (WA) type.
3. **Extract the amplitude:** Extract the maximum amplitude (zero to peak; in mm) from either the horizontal or vertical component of the synthetic WA seismogram. This should be done within a window that includes the S and surface waves.
4. **Calculation of $-\log_{10}(A_0)$:** Calculation of $-\log_{10}(A_0)$ based on the geometric spreading, anelastic attenuation, and hypocentral distance (R) of the recording station to the earthquake epicenter.

5. **M_L Calculation:** Finally, the M_L formula (Equation 2.1) can be used.

[Uhrhammer and Collins \[1990\]](#) work have demonstrated notable differences in the static magnification and damping ratio of Wood-Anderson (WA) instruments compared to the values previously reported by Anderson and Wood ([Anderson and Wood, 1925](#)), as well as those adopted by Richter ([Richter, 1935](#)). [Uhrhammer and Collins \[1990\]](#) results indicate that the static magnification is 2080 and the damping ratio is 0.7, in contrast to the values of 2800 and 0.8 reported by Anderson and Wood ([Anderson and Wood, 1925](#)) and subsequently used by Richter ([Richter, 1935](#)). This discrepancy in the original static magnification of 2800 leads to a systematic difference in the estimated magnitude, averaging 0.13 units. This underscores the importance of accurate instrument characteristics in seismic measurements, as even small errors in static magnification can have a measurable impact on the precision of earthquake magnitude estimates. ([Uhrhammer and Collins, 1990](#); [Uhrhammer et al., 1996](#)) research thus highlights the importance of refining and updating instrumental parameters for more accurate seismic analysis and reporting. As seismic technology continues to advance, a wider range of instruments has become available, such as Broadband stations and other types of short-period seismometers. In order to apply Richter's magnitude formula for other types of seismometre rather than WA, we must simulate record the Wood-Anderson (WA) seismogram. This includes synthesizing WA records by deconvolving the instrument response from digitally recorded waveforms and then convolving with the WA response. The WA response's specific characteristics significantly impact this process. The comparison of the originally recorded seismogram with the WA response convolution version is illustrated in Figure 3.12, offering a visual representation of the effect of this process on the seismic data.

The Wood-Anderson (WA) Seismograph: The Wood-Anderson (WA) torsion seismometer (Figure 2.2), which Richter used to measure maximum amplitude, became popular in 1925 after Anderson and Wood published information on it. This instrument's rapid adoption in California, as pointed out by [Uhrhammer and Collins, 1990](#), established it as the preeminent seismometer for a long time. Significantly, specific WA devices continued to function well into the 1990s. The Wood-Anderson seismometer is a short-period instrument with specific parameters. It has a static amplification of 2800, a free period of 0.8 seconds, and a damping factor of 0.8. The instrument is sensitive only to ground motion displacements above approximately 1 Hz, and frequencies below this threshold experience a rapid decrease in sensitivity. The corner frequency of the seismometer

is 1.25 Hz, effectively turning it into a high-pass filter affecting the recorded displacement. The Wood-Anderson seismometer's scale saturates during large earthquakes and shows a distinct response pattern. This occurrence primarily results from the long-period seismic energy in larger earthquakes. The duration of seismic activity correlates with the dominance of long-period energy, which typically increases with the magnitude of the earthquake. According to [Shearer \[2019\]](#), the response of the Wood-Anderson seismometer decreases the importance of long-period seismic energy, leading to the scale's saturation for major seismic events. Ground motion frequencies decrease gradually as the magnitudes of seismic events increase. This phenomenon occurs because frequencies above the high-pass filter threshold of the seismometer decrease in prominence as magnitudes exceed 7. Additionally, larger seismic events cause a significant shift in the relationship between earthquake size and radiated S-wave characteristics. The transformation is influenced by finite fault effects, which result in less indicative radiated S-waves of the entire slip. S-waves are emitted from multiple sub-faults dispersed throughout the whole rupture zone during significant earthquakes. The complex distribution poses challenges to establishing a correlation between the S-wave amplitude (A) and the actual size of the seismic source. Hence, the measured motion does not demonstrate the same accelerated amplification rate characteristic of smaller seismic events. [Uhrhammer and Collins \[1990\]](#) work have demonstrated notable differences in the static magnification and damping ratio of Wood-Anderson (WA) instruments compared to the values previously reported by Anderson and Wood, as well as those adopted by Richter. [Uhrhammer and Collins \[1990\]](#) results indicate that the static magnification is 2080 and the damping ratio is 0.7, in contrast to the values of 2800 and 0.8 reported by Anderson and Wood and subsequently used by Richter. This discrepancy in the original static magnification of 2800 leads to a systematic difference in the estimated magnitude, averaging 0.13 units. This underscores the importance of accurate instrument characteristics in seismic measurements, as even small errors in static magnification can have a measurable impact on the precision of earthquake magnitude estimates. [Uhrhammer and Collins, 1990](#) research thus highlights the importance of refining and updating instrumental parameters for more accurate seismic analysis and reporting.

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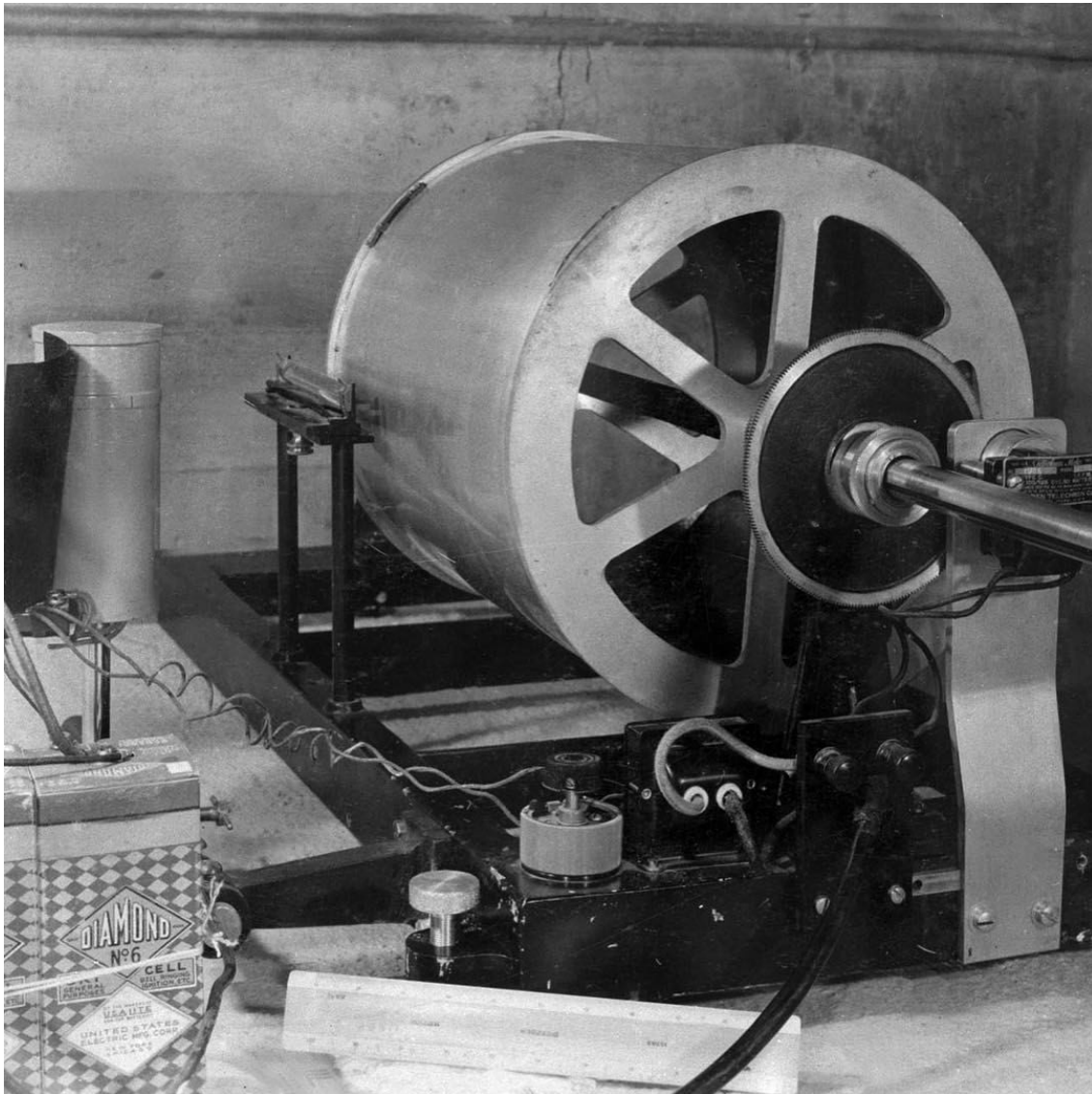


Figure 2.2: The Wood-Anderson Seismograph – An Early Instrument for Measuring Earthquake Magnitude.

record the Wood-Anderson (WA) seismogram. This includes synthesizing WA records by deconvolving the instrument response from digitally recorded waveforms and then convolving with the WA response. The WA response's specific characteristics significantly impact this process, as shown in Figure 1 with a corner frequency of 1.25 Hz.

2.2.2 Duration Magnitude (M_d)

The duration magnitude (M_d) is a commonly used measure for local earthquakes. It is widely accepted in numerous regional networks due to its simplicity and reliability in estimating earthquake size. This is achieved through a simple procedure centered on measuring the duration of recorded seismograms. In 1958, [Bisztricsany, 1958](#) established a correlation between magnitude and duration. Since then, several authors, such as ([Sole'vov, 1965](#); [Tsumura, 1967](#); [Bakun, 1984](#); [Vidal and Munguía, 2005](#)), have examined the duration of recorded seismograms as an indicator of event size. The duration magnitude depends on the duration of ground shaking, hypocentral distance, and a station correction coefficient. [Real and Teng, 1973](#) and [Herrmann, 1975](#) formally define duration magnitude as follows:

$$M_d = a + b * \log(\tau) + c * R + S \quad (2.2)$$

The equation includes variables for signal duration (represented by τ), hypocentral distance (represented by R), and station correction (represented by S). The coefficients a , b , and c are determined through regression analysis. Several methodologies have been proposed for determining the duration of a signal or coda such as:

1. The duration is measured from the beginning of the P-wave to the end of the coda, where the signal merges with the seismic noise of the same frequency ([Herrmann, 1975](#)).
2. The duration is computed from the onset of the P-wave to the point where coda amplitudes have attenuated to a specific threshold level. This threshold can be defined as the average signal-to-noise ratio, absolute signal amplitudes, or signal level.
3. The total elapsed time is calculated by subtracting the coda threshold time from the origin time of the event.

2.2.3 Surface Wave Magnitude (M_s)

Gutenberg expanded Richter's concept of earthquake magnitude to accommodate larger events by focusing on longer-period seismographs and the measurement of surface waves. Specifically, he developed the first surface-wave magnitude scale in 1945 ([Gutenberg, 1945a](#)), based on teleseismic recordings of surface waves traveling along oceanic paths at distances greater than 20° . Gutenberg's formula suggested measuring the maximum amplitude ($A_{\max}$) in the surface-wave train for periods around 20 seconds within a distance range of 20° to 160° . For distances over 200 km ([Havskov and Ottemoller, 2010](#)), broadband seismograms and long-period seismograms (especially for distances exceeding 600–1,000 km) captured surface waves with periods from about 3 to 60 seconds, typically around 20 seconds, which became a key in this magnitude scale.

This scale became a standard for estimating the magnitude of large earthquakes, providing values aligned with local magnitude (M_L) estimates and widely used by the USGS and media until the introduction of the moment magnitude scale (M_w). Despite being largely replaced by M_w for its greater accuracy in energy estimation, this amplitude-based magnitude is still often the first reported for large earthquakes due to its quicker calculation. It is also useful in distinguishing between natural earthquakes and underground explosions.

2.2.4 Body Wave Magnitude (m_b)

The body wave magnitude scale (m_b) is one of the most widely used scales for measuring earthquakes recorded at teleseismic distances. This magnitude is calculated from the maximum amplitude within the P-wave train recorded by a short-period seismograph, up to the phase PP. The m_b scale evolved from the original body wave magnitude developed by ([Gutenberg, 1945b](#); [Gutenberg and Richter, 1945](#)), who used instruments with broader response capabilities than short-period seismographs. Since different response curves exist for short-period seismographs, standardization is essential.

Unlike the local magnitude (M_L), which is measured on a simulated Wood-Anderson seismogram without requiring gain correction, m_b measures the maximum ground displacement directly. The m_b scale is defined as:

$$m_b = \log \left(\frac{A}{T} \right) + Q(\Delta, h)$$

where A is the ground displacement in microns, T is the dominant period of the measured waves, Δ is the epicentral distance in degrees, and Q is an empirical function that adjusts for distance and event depth (e.g., [Veith and Clawson, 1972](#)). This Q function incorporates details of average amplitude variations with epicentral distance and source depth.

Typically, m_b is measured from the first few cycles of the P-wave arrival on short-period vertical-component instruments, where the dominant wave period is around 1 second. As with local magnitude, m_b estimates for the same event vary between stations, with differences up to ± 0.3 due to factor, like radiation pattern, directivity, and local site effects. To adjust for stations with consistently high or low m_b values, station correction terms are often applied.

The optimal range for epicentral distances is 20–100°. The attenuation function recommended by IASPEI is the original function developed by [Gutenberg and Richter, 1956](#) for P-waves recorded on vertical seismographs (see [Figure 2.3](#)). This function is currently used by both ISC, CRAAG, and USGS.

Since m_b measures amplitude at a frequency near 1 Hz, it saturates below magnitude 7. An alternative, m_B , can be derived using unfiltered broadband velocity records and the Gutenberg-Richter attenuation relation, which increases the saturation threshold compared to m_b . Although m_B is not constrained by specific instrument responses and thus could be a more versatile magnitude scale, it is rarely used and is not routinely reported by USGS or ISC. A limitation of both m_b and m_B is that they tend to underestimate magnitude when compared to the surface wave magnitude (M_s).

2.2.5 Moment Magnitude (M_w)

After [Aki, 1966](#) introduced the seismic moment M_0 (measured in Newton.meters), earthquake 'size' assessment gained prominence through the moment magnitude (M_w) scale proposed by [Kanamori and Stewart, 1977](#) and further developed by [Hanks and Kanamori, 1979](#). M_w represents a modern magnitude scale that is directly linked to the well-defined physical parameter of the seismic moment, M_0 . The relationship between moment magnitude (M_w) and seismic moment (M_0) is described by the [Equation 2.3](#), where M_w is the moment magnitude, and M_0 is the seismic

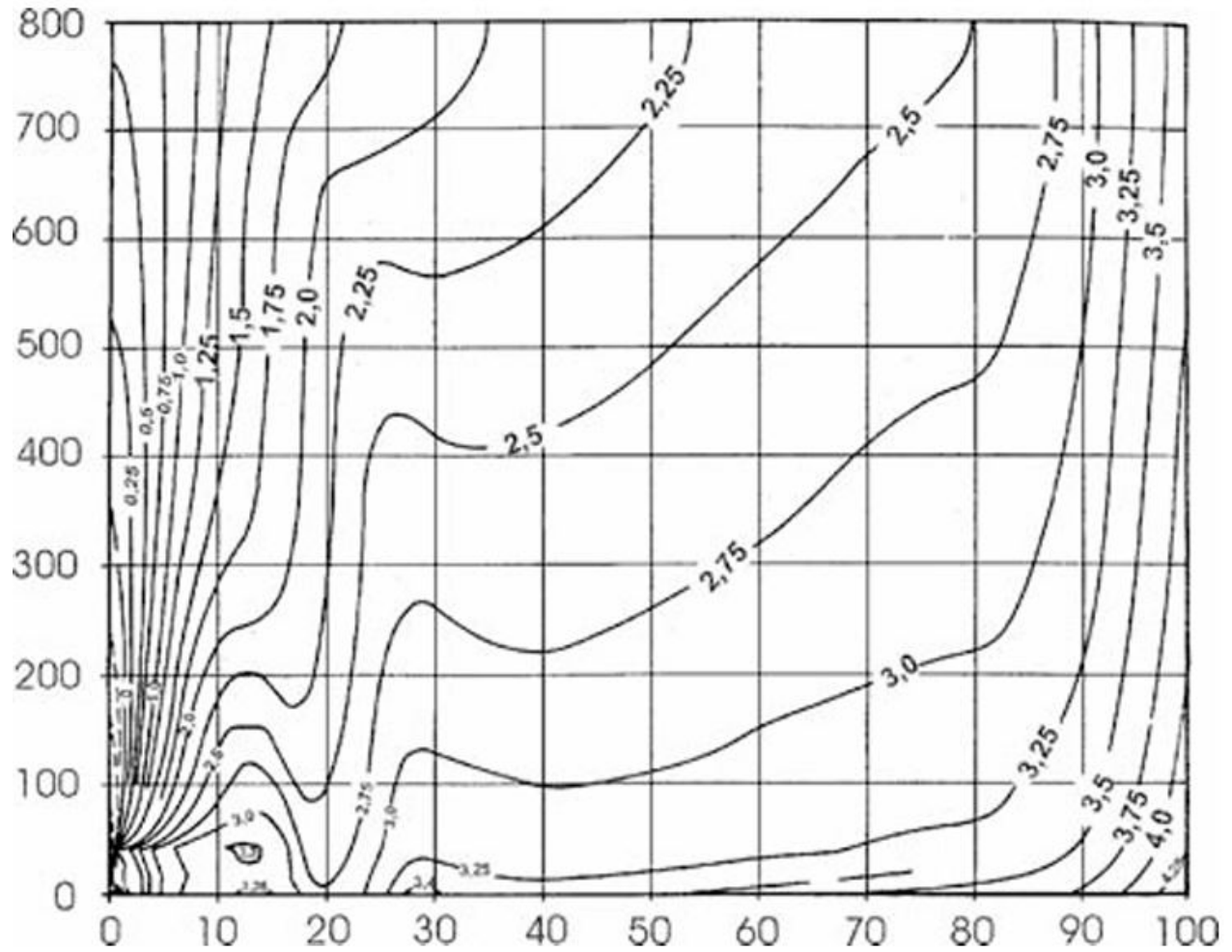


Figure 2.3: Attenuation functions $Q(\Delta, h)$ for m_b determination from short-period, vertical component records typical of WWSSN instruments (Veith and Clawson, 1972). The x-axis represents the epicentral distance in degrees, and the y-axis represents the depth.

moment measured in N.m.).

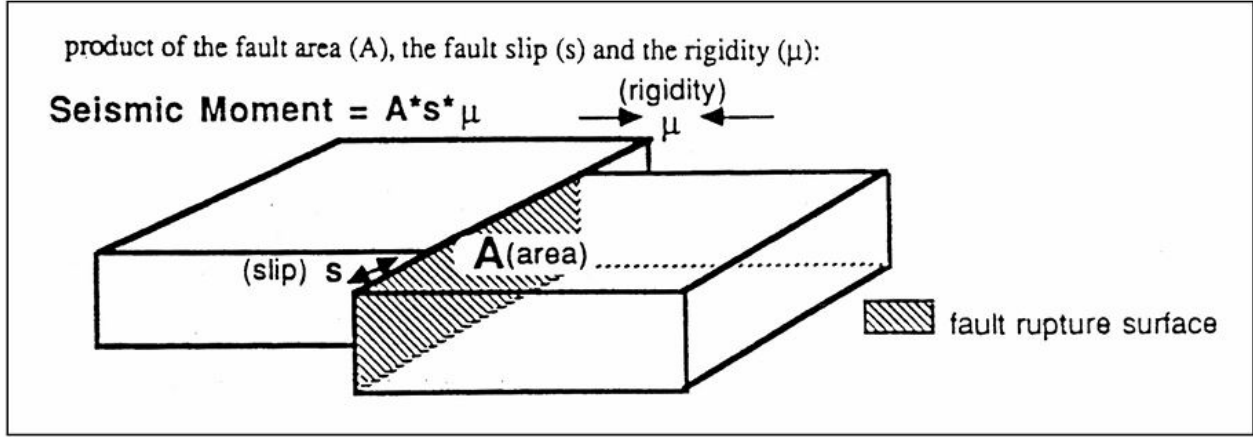


Figure 2.4: The seismic moment reflects the physical properties of an earthquake source. It is calculated as the product of the fault rupture area (A), the slip or displacement along the fault (s), and the rigidity (μ), which represents the stiffness of the fault material.

M_w , essentially a representation of M_0 , is commonly preferred over M_0 due to its convenience and natural scale, as pointed out by [Kanamori, 1983](#). It is directly connected to physical source properties through M_0 , enabling its inference even for historical seismic events based on estimated rupture sizes. Unlike other magnitude measures, M_w does not saturate, as highlighted by [Kanamori and Stewart, 1977](#). This unique characteristic makes it a preferred base of comparison for other magnitude scales. As the determination of M_0 relies on physics-based models, rather than simple empirical relations, obtaining accurate values can be challenging ([Kanamori, 1983](#)). Various strategies have been designed to address this difficulty. The moment magnitude is calculated using the following relationship:

$$M_w = \frac{2}{3} \log_{10}(M_0) - 10.7 \quad (2.3)$$

Where:

- M_0 : Seismic moment, typically expressed in dyne · cm,
- 10.7: A constant that normalizes the scale to align with traditional magnitude values for smaller earthquakes.

The seismic moment (M_0) is derived from the physical properties of the fault and the seismic source. For a homogeneous and isotropic medium ([Figure 2.4](#), [Aki, 1966](#) provided the following formula:

$$M_0 = A * S * \mu. \quad (2.4)$$

Where:

- μ : Rigidity or shear modulus of the medium (3×10^{11} dyne/cm² for crustal rocks),
- S : Average final displacement after the rupture,
- A : Surface area of the rupture.

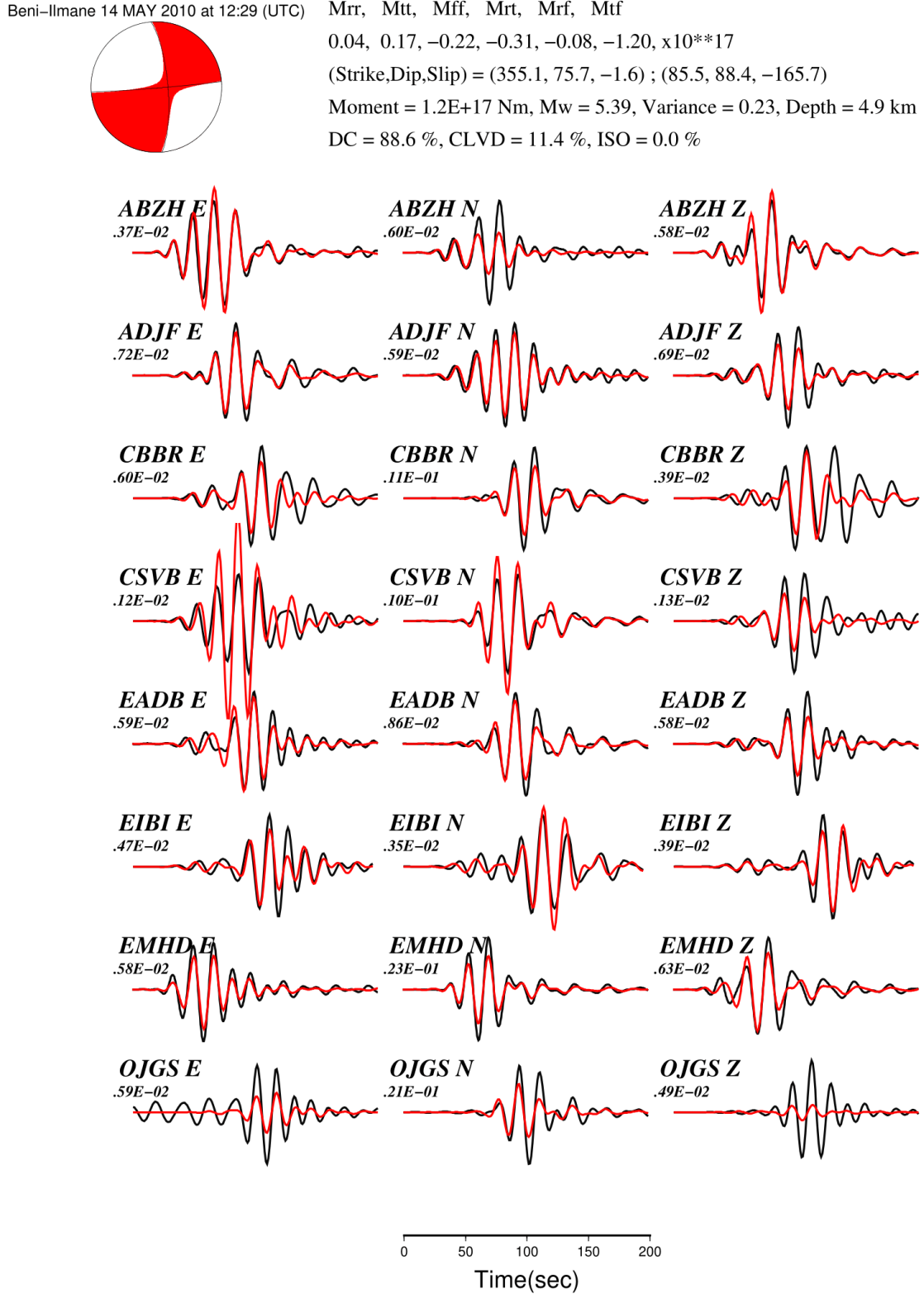


Figure 2.5: Waveform inversion process for determining the seismic moment (M_0): The observed and synthetic waveforms are compared to extract source parameters. This method incorporates the effects of fault geometry, source mechanisms, and propagation paths to estimate M_0 .

Alternatively, the seismic moment can be estimated from the spectral characteristics (Figure 2.6) of low-frequency seismic waves using:

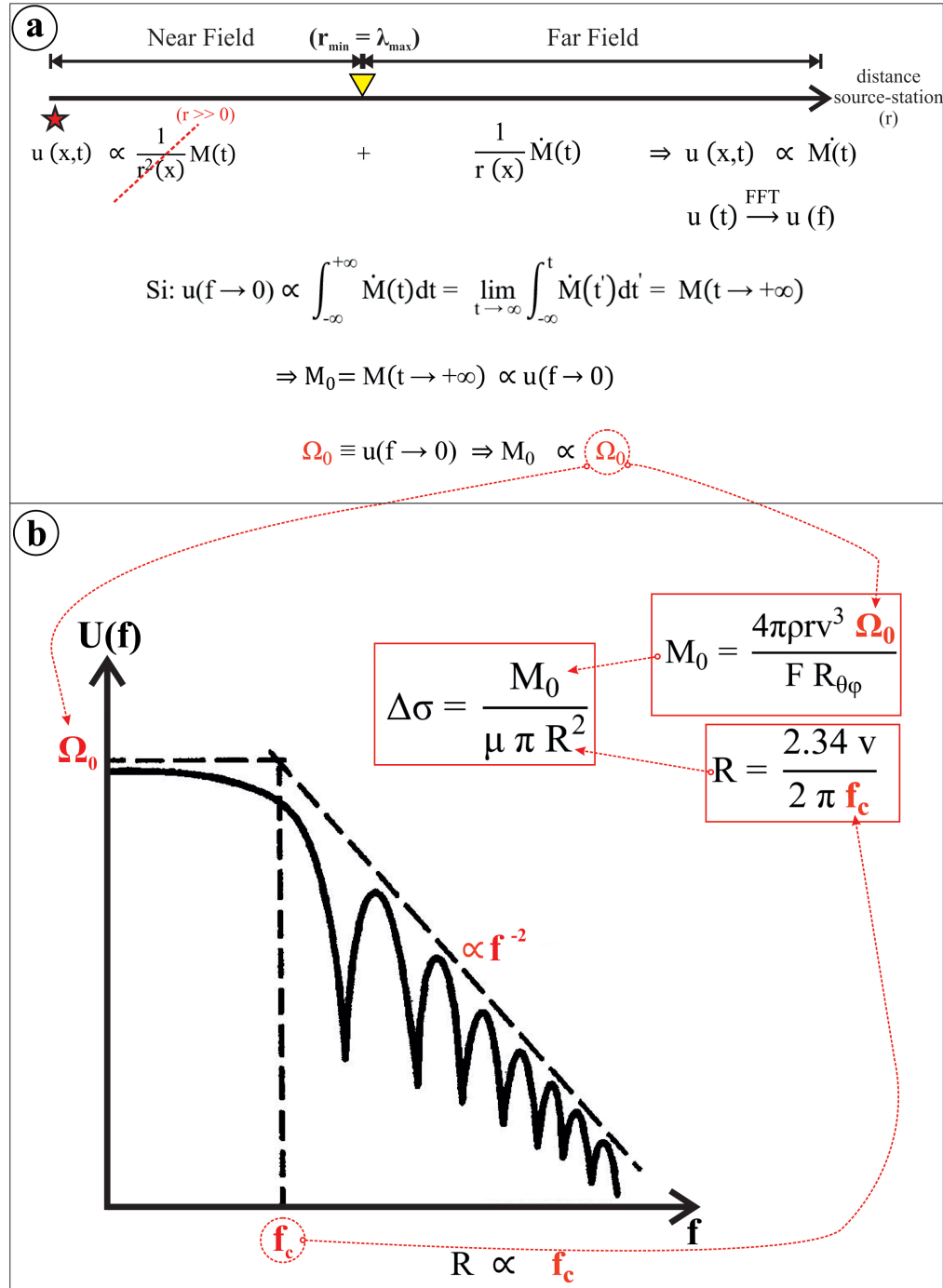


Figure 2.6: Spectral analysis of seismic waves: This Figure illustrates the amplitude spectrum of seismic waves (P and S) recorded at a specific station. The low-frequency plateau is proportional to the seismic moment (M_0), while the corner frequency (f_c) and high-frequency decay provide insights into source properties such as fault size and stress drop.

$$M_0 = \frac{4\pi\rho v^3 R\Omega_0}{F \cdot R_{(\theta,\phi)}} \quad (2.5)$$

where:

- M_0 : Seismic moment (in $\text{N} \cdot \text{m}$),
- ρ : Density of the medium near the source (in kg/m^3),
- v : (P or S)wave velocity (in m/s),
- R : Hypocentral distance (in m),
- Ω_0 : Low-frequency spectral amplitude,
- F : Geometric spreading factor,
- $R_{(\theta,\phi)}$: Radiation pattern correction factor, which depends on the azimuth (θ) and take-off angle (ϕ).

The moment magnitude scale addresses limitations of traditional magnitude scales by providing a uniform measure of earthquake size across all magnitudes. Unlike scales such as the Richter magnitude, which saturates for larger earthquakes, M_w accurately captures the energy released by both small and large seismic events. This makes M_w a critical tool for assessing earthquake impacts, modeling seismic hazard, and understanding tectonic processes.

2.3 Why are different magnitude values announced by different agencies when an earthquake occurs?

Many people are familiar with the term "Richter Scale" and often associate it with earthquake magnitude. However, this understanding can be misleading, as multiple magnitude scales exist, each quantifying earthquake size in different ways. In Algeria, for instance, the media frequently refers to all seismic magnitudes as "The Richter Scale," despite the unique characteristics of each scale. This oversimplification can lead to confusion.

The concept of measuring earthquake size using a logarithmic scale originated with Charles Richter in the 1930s. His original scale, local magnitude (M_L), was designed to assess earthquakes in Southern California, utilizing high-frequency seismic data from nearby stations. As seismic monitoring expanded globally, it became clear that Richter's method was only accurate within certain frequency and distance ranges. To address this limitation, seismologists introduced new scales, such as body wave magnitude (m_b), duration magnitude (M_d), and surface wave magnitude (M_s). Each scale operates within a specific frequency range, capturing distinct aspects of seismic activity, but they align closely with the Richter scale within their valid ranges.

With these different scales, it's not uncommon for multiple magnitude values to be reported for the same earthquake, sometimes varying by more than a half a unit. In the 1960s and 1970s, a notable issue was that most commonly used magnitude scales "saturated" for large earthquakes. For example, the 1906 San Francisco earthquake was assigned a surface wave magnitude (M_s) of 8.3, the 1960 Chile earthquake an M_s of 8.5, and the 1964 Alaska earthquake an M_s of 8.4, suggesting similar sizes. However, these events had vastly different fault rupture areas and durations, indicating that M_s alone could not fully capture their differences (as shown in Table 2.1).

Earthquake	Area of fault repture	Duration	M_s	M_w
1906 San Francisco	5000 km^2	40 seconds	8.3	7.8
1960 Chile	100000 km^2	07 minutes	8.5	9.5
1964 Alaska	50000 km^2	05 minutes	8.4	9.2

Table 2.1: Magnitudes and fault dimensions.

Earthquakes also exhibit complex wave patterns due to variations in fault orientation and rupture directivity, which can direct seismic waves more strongly in certain directions. As a result, networks with different instrumentation and geographic coverage, such as the ADSN in Northern Algeria and the National Earthquake Information Center's global network, may report slightly different magnitudes.

For instance, the May 14, 2010, Beni Ilmane earthquake (Figure 2.7), approximately 300 km

southeast of Algiers, highlights these variations. Here's how different agencies assessed its magnitude:

1. The Centre for Research in Astronomy, Astrophysics, and Geophysics (CRAAG) released an initial duration magnitude (M_d) estimate of 5.2.
2. The Laboratoire de Détection et de Géophysique (LDG) issued a local magnitude (M_L) estimate of 5.0.
3. The International Seismology Centre (ISC) established a surface wave magnitude (M_s) of 4.9.
4. The Instituto Andaluz de Geofísica (IAG) estimated the moment magnitude (M_w) to be 5.4.
5. The Centre Sismologique Euro-Méditerranéen (CSEM) determined a surface wave magnitude (M_s) of 4.9

These different estimates, often collectively labeled "Richter Magnitude" by the media, were not calculated in Richter's original manner. However, all scales (except M_w) are generally calibrated so that a magnitude 5 event on any scale is roughly equivalent, typically indicating a moderate earthquake capable of causing minor local damage.

The m_b and M_s scales were initially developed to align with the M_L scale for smaller, local events in California. However, it is not feasible to make these scales match consistently across all event sizes due to differences in their measurement periods and the way earthquake frequency content varies with event size. For larger events, the source spectrum typically decreases as f^{-2} beyond a certain corner frequency, f_c , which shifts to lower frequencies as event size increases.

Assuming constant stress drop, the fault dimension and rupture duration scale approximately with $M_0^{1/3}$, where M_0 represents the seismic moment. As a result, the corner frequency is inversely proportional to rupture duration and scales as $M_0^{-1/3}$, causing the position of f_c to move to lower frequencies as events become larger (Figure 2.8).

$$f_c \propto M_0^{-1/3}$$

At frequencies below f_c , a linear relationship exists between magnitude (specifically, the $\log_{10}$ of the measured amplitude) and seismic moment. However, at higher frequencies, this linearity

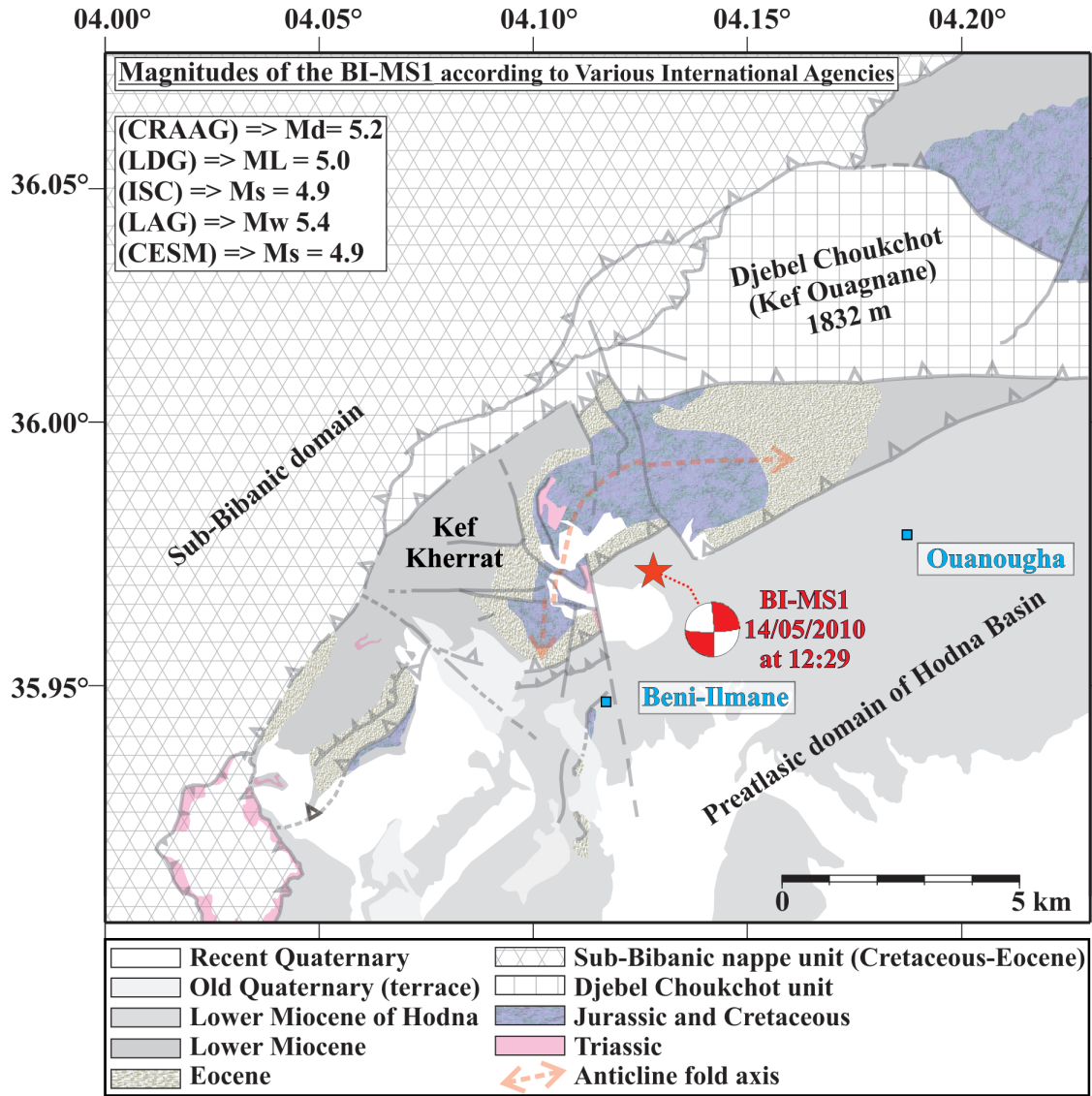


Figure 2.7: Magnitude Estimates of the Beni Ilmane Earthquake from Various Agencies.

diminishes, and the magnitude scale fails to fully account for increasing event sizes. This phenomenon is known as **magnitude saturation**. At a given measurement frequency (e.g., 1 Hz for m_b), saturation occurs when the event size becomes large enough for the corner frequency to fall below the measurement frequency. Although not all earthquake spectra fall off exactly at f^{-2} , any degree of falloff will contribute to some level of magnitude saturation.

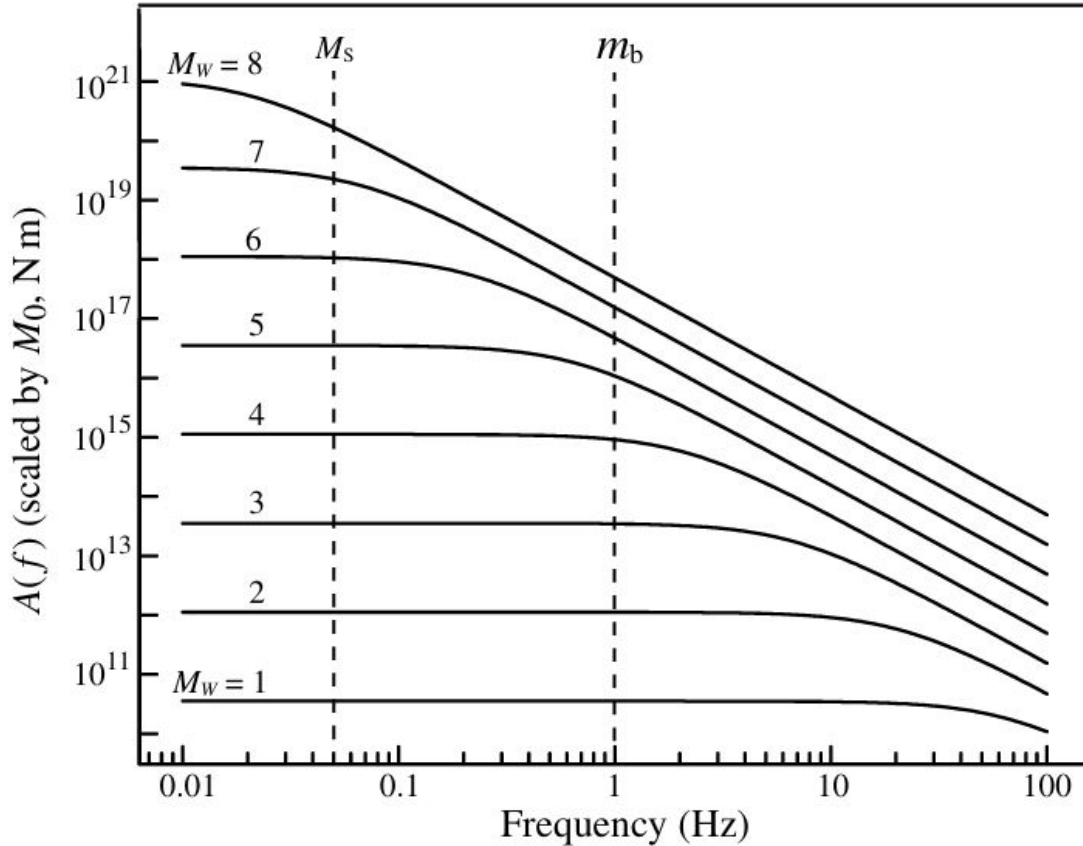


Figure 2.8: For larger events, the corner frequency in the source spectrum shifts to lower values, limiting the observed increase in amplitude at the fixed frequencies typically used for estimating M_s (20s) and m_b (1s).

Another factor contributing to magnitude saturation is the fixed window length used to measure amplitudes. This window may be too short to capture the complete amplitude of larger events, as observed by [Houston and Kanamori, 1986](#). In practice, m_b values start to saturate around $m_b = 6.5$, while M_s values, measured over longer periods, saturate around $M_s = 7.5$. Consequently, it is rare for m_b to exceed 7 or M_s to exceed 8.5, even for exceptionally large earthquakes.

Saturation occurs because seismic wave amplitudes reach a peak that doesn't increase propor-

tionally with earthquake size after a certain point. For instance, body wave magnitude (m_b) saturates around 6.5- 6.8, while surface wave magnitudes cap around 8.3. In response, seismologists developed the moment magnitude scale (M_w) in the 1970s. By focusing on the physical parameters of the fault—such as rupture area, average displacement, and rock rigidity—this scale provides a more accurate measure of the energy released, especially for large earthquakes.

Various seismic magnitude scales have become integral to the field of seismology due to their historical and practical significance. However, when quantifying earthquake size with a single value, it is more accurate to use seismic moment, as it directly relates to a fundamental physical property of the source. Moment also lends itself to geodetic studies and allows for meaningful comparisons with long-term geological slip rates.

While it is possible to interpret the average behavior of different magnitude scales relative to moment, it is essential to recognize that measurements like M_L , m_b , and M_s often exhibit significant variability among events with the same seismic moment. This variability is likely not random; rather, it may reflect differences in stress drop or other source characteristics, with systematic regional variations observed in some cases ([Ekström and Dziewonski, 1988](#)).

2.4 Conclusion

In conclusion, the various magnitude scales discussed throughout this chapter are fundamental tools for understanding and quantifying earthquake size, each serving specific purposes based on the characteristics of the seismic event and the available data. While moment magnitude (M_w) has emerged as the most comprehensive scale for large earthquakes, it is essential to acknowledge the continued importance of local magnitude (M_L) and duration magnitude (M_d) in certain contexts, especially for smaller, regional, and local events.

Local magnitude (M_L) is especially valuable for estimating the size of earthquakes in regions with dense seismic networks and where high-frequency seismic data are available. Despite its saturation for large events, M_L provides an accessible and efficient means of assessing moderate to small earthquakes. It is particularly useful in local contexts, such as northern Algeria, where quick assessments of earthquake size are often needed for early warning and response. Its long-standing application in seismic practice means it remains a standard method for local seismic hazard studies.

Duration magnitude (M_d), on the other hand, plays a crucial role in capturing the total energy released over time, especially for earthquakes with extended durations or those occurring in deeper regions. M_d can often be more accurate for large, distant, or deep events where other scales, like M_L , may fail to provide an accurate representation of the earthquake's size. By focusing on the duration of seismic shaking, M_d complements traditional scales and provides additional insights into the nature of the event.

The differences in the application and reliability of these magnitude scales underscore the necessity of using multiple methods to assess earthquake size, particularly when dealing with complex seismic scenarios. While M_w offers a comprehensive and reliable measure for large events, M_L and M_d remain invaluable, especially for local, moderate, and distant earthquakes, where they can provide essential data for understanding seismicity and its potential impacts. The careful selection of the appropriate magnitude scale allows seismologists to more accurately model seismic hazard, improve earthquake prediction, and enhance earthquake preparedness.

Chapter 3

Data and Methods

This chapter provides a detailed overview of the data acquisition and analysis methods employed in this thesis. The study is based on data from the Algerian Digital Seismic Network, which serves as the cornerstone of our research and underpins all subsequent analyses. The chapter begins by outlining the network's development, highlighting advancements in infrastructure, technology, and instrumentation. Key improvements include the deployment of advanced seismic sensors, the adoption of diverse data transmission techniques, and the establishment of robust quality control systems each contributing significantly to the accuracy and reliability of the seismic data.

Additionally, the chapter discusses the operational aspects of the network, such as monitoring, data backup protocols, and data processing techniques, which collectively contribute to the integrity and precision of the seismic data used throughout this study. These elements are essential for maintaining high-quality data, which is the cornerstone of the seismological analyses presented in this thesis.

In the following sections, we outline the systematic stages of data processing utilized in this work. These stages are part of standardized procedures followed in analyzing multiple seismic sequences, as demonstrated in several studies co-authored by the author, including those focused on regions such as Beni Ilmane ([Abacha et al., 2023a](#)), El-Aouana ([Abacha et al., 2023b](#)), and Jijel ([Yelles-Chaouche et al., 2021](#)). The data preparation workflow is discussed in detail, starting with the installation of seismic stations and retrieval of raw data. Key preprocessing steps, such as time corrections, deconvolution, and filtering, are explained, culminating in the organization of the data into a coherent and structured database for subsequent analysis. Lastly, the chapter discusses

the determination of seismic wave arrival times, which is essential for accurately defining the time window used in the estimation of local and duration magnitudes, as well as other critical parameters for earthquake characterization.

This chapter lays the groundwork for the methodological framework of this thesis, ensuring that the seismic data are processed and interpreted in a consistent and reliable manner, ultimately supporting the study's conclusions.

3.1 Data and networks

3.1.1 Algerian seismological networks

1- Early Instrumentation and the Analog Network

The history of instrumental seismology in Algeria began in 1910 with the installation of the first mechanical seismograph, a Mainka-Bosch model, at the Bouzareah Observatory (Roussel, 1973). Over the following years, additional seismic stations equipped with Grenet-Coulomb short-period seismographs were established in various towns across northern Algeria. After the devastating 1980 El Asnam earthquake (M_S 7.3), Algeria secured funding in 1985 to upgrade and modernize its seismic monitoring system. By 1990, a new seismological network, known as REALSAS (Réseau ALgérien de Surveillance et d'Alerte Sismique) or the Telemetered Network, was established. This network consisted of 32 stations using radio link technology, distributed across four regional zones: Algiers, Constantine, Chlef, and Oran (Bezzeghoud et al., 1994). At this time, four stations, referred to as fixed stations, remain operational (Figure 3.1). Unfortunately, due to a major political-security crisis in the country during the 1990s, the network was deactivated after only one year of operation, causing seismological research to lag significantly and limiting it to basic descriptions of seismic events. Although the network was partially restored in 1998, it wasn't until the 2003 Boumerdes earthquake (M_w 6.9) that substantial efforts were made to enhance Algeria's seismological monitoring capabilities.

2- The Digital Network

The most crucial components of a seismological network are the sensors, or seismometers, which are categorized according to their sensitivity to specific frequency ranges. Each sensor func-

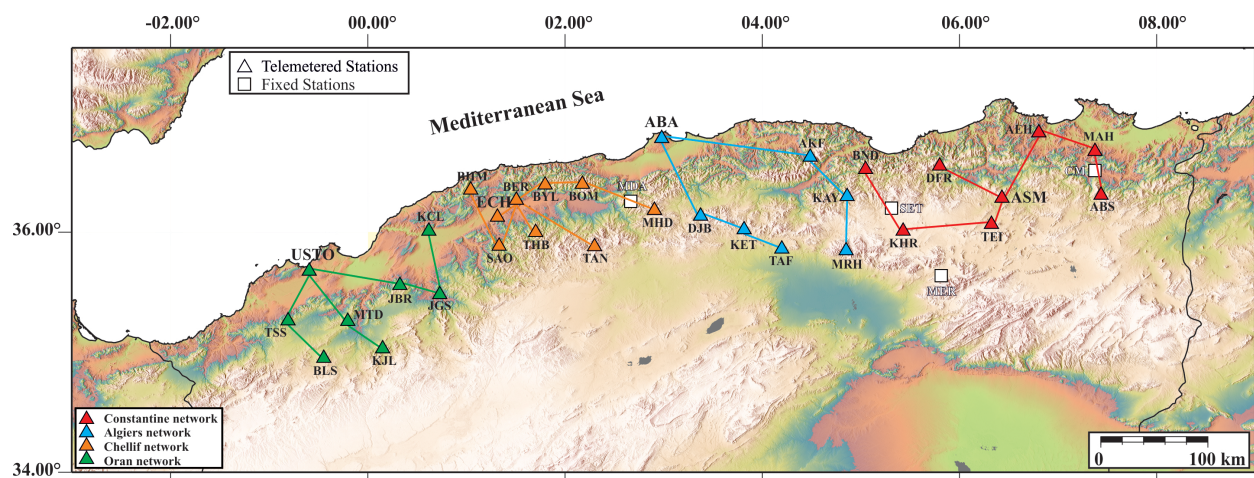


Figure 3.1: The Algerian REALSAS (Réseau ALgérien de Surveillance et d'Alerte Sismique) Network or Telemetered Network. Station colors represent the four regional networks: Constantine (East), Algiers (Central), Chellif (Central-West), and Oran (West).

tions as an oscillator with a limited bandwidth tied to its natural frequency, meaning it cannot respond to all frequencies within a seismic signal. For this reason, the Algerian Digital Seismic Network (ADSN, [Yelles-Chaouche et al., 2013](#)) includes a variety of seismometers designed to cover different frequency ranges, such as "short-period," "broadband," and "very broadband" seismometers. Since 2006, a network featuring 10 broadband stations and 40 short-period stations has been progressively deployed across northern Algeria. Presently, the Algerian Digital Seismic Network (ADSN) encompasses 72 permanent stations including 20 broadband (three-component) and 52 mono-component short-period ([Yelles-Chaouche et al., 2013](#); [Yelles-Chaouche et al., 2022](#)). These permanent stations are equipped with Geodevice BBVS-60s and Streckeisen STS-2 broadband sensors, linked to Geodevice EDAS-24IP and Q330 digitizers, respectively. Additionally, the short-period stations are equipped with 1-Hz SS-1 sensors coupled to Q330 digitizers (Figure 3.2a). All these stations are controlled first by Antelope systeme (2006-2020) and since 2021, by both Antelope and Seiscomp systems., which ensure real-time connectivity through a variety of transmission means, including satellite, internet and cell phones. The establishment of a high-performance digital seismological network was a significant transition in the development of seismological studies in the country for the comprehensive assessment of seismic hazards.

3- The Temporary Network

In the event of a major earthquake, a temporary network of portable stations is rapidly deployed

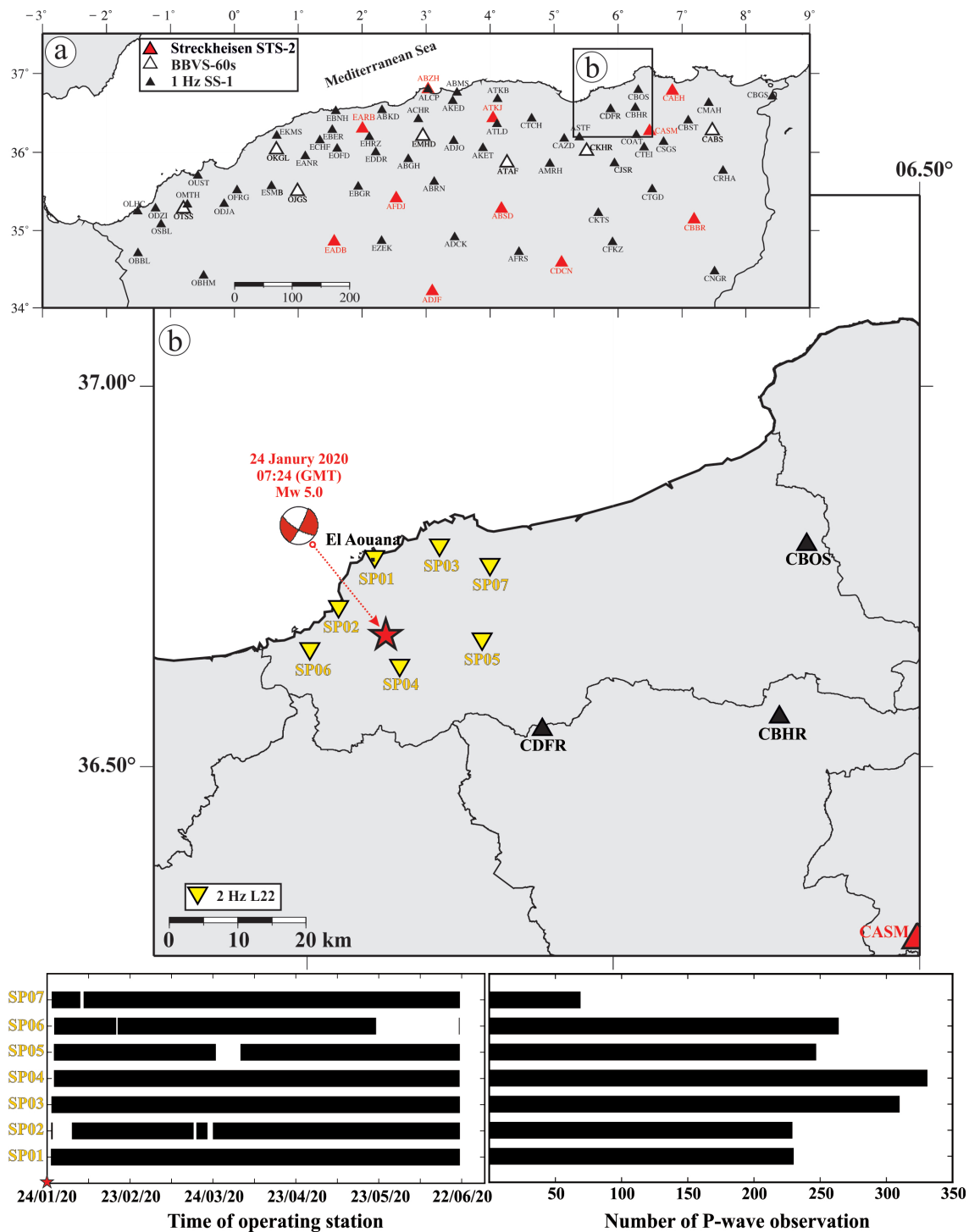


Figure 3.2: Example of the data acquisition during the 2020 El Aouana Earthquake (modified from [Abacha et al., 2023b](#)). (a) Permanent ADSN seismic stations. (b) The seven portable stations of the temporary network. The station operation timeline and the number of P-phase pickings for each station are displayed in the bottom left and right corners, respectively.

in the epicentral area to capture the maximum number of aftershocks, including those of low magnitude, and to ensure good azimuthal coverage. Figure 3.2b illustrates a field campaign following the 2020 El Aouana earthquake (Abacha et al., 2023b), in which I actively participated. In this example, seven stations were installed, each equipped with Omnirecs CUBE3 24-bit digitizers and Mark Products L22 three-component sensors (fc = 2.0 Hz), supplemented with solar panels, and configured to operate in continuous acquisition mode at a sampling rate of 100 Hz (Figure 3.3). Additionally, other types of stations are used for different seismic sequences, including Geodevice EDAS-24IP digitizers paired with 1-Hz Geodevice FSS-3 sensors, Q330 digitizers paired with 1-Hz SS-1 sensors, and SARA SL-06 24-bit digitizers coupled with SARA SS-1 sensors.

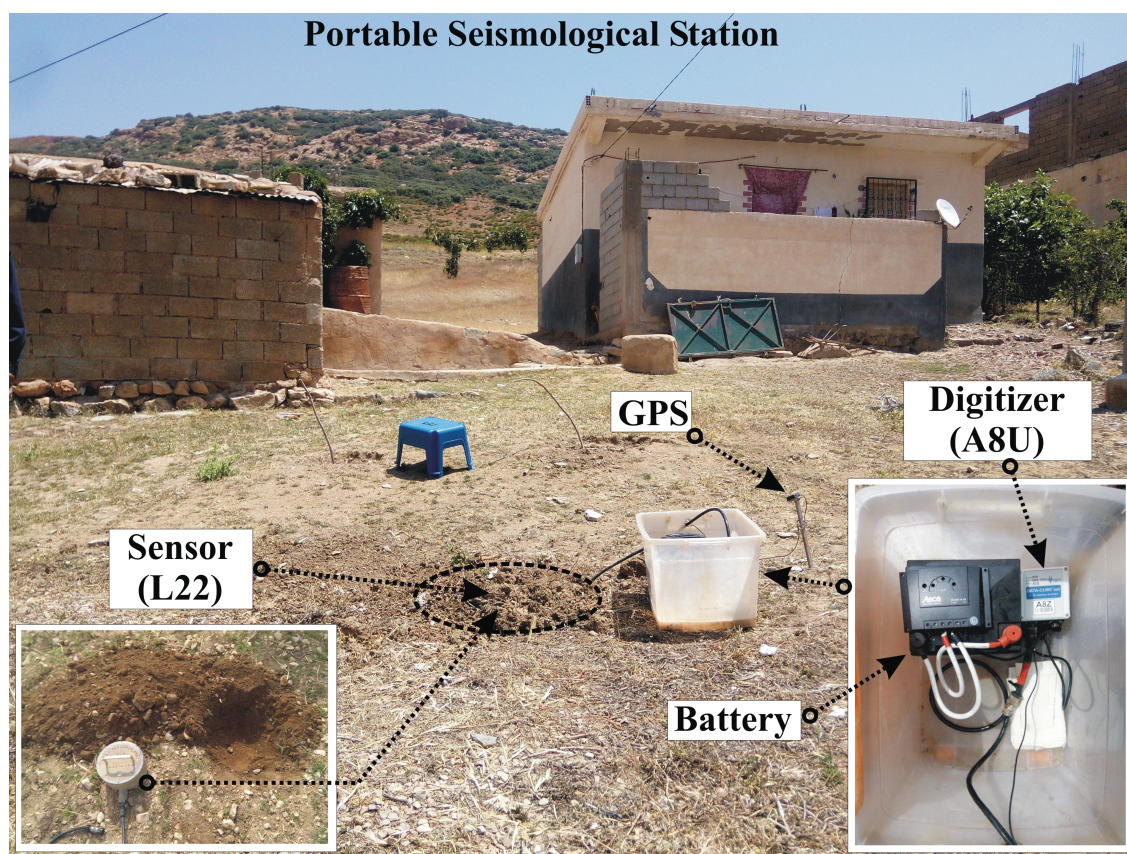


Figure 3.3: Components and Installation of a Portable Seismological Station.

3.1.2 Characterization of Seismological Data and Used Instrumentation

This section highlights the primary seismic sequences I contributed to during the data preparation process, utilizing data recorded in regions such as Béjaia-Babors, Beni Ilmane, and Mila, among

others. The recordings were obtained using both temporary local networks and the permanent Algerian Digital Seismic Network (ADSN). The dataset primarily comprises eight seismic aftershock sequences, along with some isolated events that occurred in northern Algeria (Figure 3.4).

1. Beni Ilmane Sequence (2010) The 2010 Beni Ilmane seismic sequence struck the region with a series of three main shocks of magnitude $M_w > 5$, followed by a substantial number of aftershocks, totaling approximately 25,000 events detected in the first week alone. Extensive research has been conducted on this sequence, including studies by (e.g., [Zazoun et al., 2012](#); [Abacha et al., 2014](#); [Yelles-Chaouche et al., 2014](#); [Beldjoudi et al., 2016](#); [Hamdache et al., 2018](#); [Beldjoudi, 2020](#); [Abacha et al., 2023a](#); [Rahmani et al., 2023](#); [Tikhamarine et al., 2024](#)).

The first fault model was proposed by [Yelles-Chaouche et al., 2014](#), based on data from 1,403 events relocated using HypoDD and arrival times. This model identified two fault structures: a left-lateral strike-slip fault oriented NNE-SSE and a reverse fault oriented approximately E-W. In contrast, the latest fault model presented by [Tikhamarine et al., 2024](#), utilizing a dataset containing more than 2800 events that is 100% larger than previous studies (~ 1400 event) (Figure 3.5a), indicates that the 2010 Beni Ilmane sequence actually activated 14 fault segments. This finding challenges earlier, oversimplified models that suggested only two or three faults were involved. The refined model, developed through advanced data processing and multiplet analysis, offers a more nuanced understanding of the fault geometry and the underlying driving mechanisms of the seismic activity in the region.

2. Béjaia-Babors Sequence (2012-2013) The second sequence, known as the Béjaia-Babors sequence, occurred between 2012 and 2013 and consisted of 345 events (Figure 3.4a). This sequence led to the cartography of a new transverse fault in the Babors region, referred to as the Babors Transverse Fault (BTF), which is oriented NW-SE and exhibits right-lateral strike-slip motion ([Boulaiahia et al., 2021](#)).

3. Hammam Melouane Sequence (2013-2016) The Hammam Melouane seismic sequence is divided into three subsequences, occurring over four consecutive years: 2013, 2014, 2015, and 2016. This thesis utilizes data from 2013 to 2015 only. In the 2013 sequence, a total of 361 events were detected, 167 of which were successfully localized. The horizontal distribution shows two parallel clusters of different sizes, oriented NE-SW, with the main shock located in the smaller cluster. In the 2014-2015 sequence, 139 events were detected, out of which 106 were successfully

localized. The horizontal distribution again reveals two clusters: one oriented NW-SE and the other NNW-SSE, with the main shock located in the larger cluster (Figure 3.5c).

4. Ain Azel Sequence (2015) From March 15 to April 30, 2015, the Ain Azel seismic sequence recorded a total of 1,008 events, which were manually identified and subsequently localized. The distribution of the 856 selected events, based on localization errors, revealed two distinct swarms: the first swarm, highlighted in blue, is oriented ENE-WSW, while the second swarm, marked in green, is oriented WNW-ESE (Figure 3.5b).

5. Jijel Sequence (2019) The fifth sequence took place in Jijel in 2019 and was characterized by a limited number of aftershocks only 14 events, including the main shock, over a span of two days. The seismic activity likely subsided quickly afterward or produced low-magnitude events that were too distant from the nearest temporary station, "SP01," to be recorded (Figure 3.4b). Tectonically, this event is likely associated with a thrust fault oriented approximately E–W, which is probably the western extension of the eastern segment (Yelles-Chaouche et al., 2021).

6. El Aouana Sequence (2020) The sixth sequence occurred in the El Aouana region in 2020 and comprised 360 events. The spatial distribution of these events aligned along a segment of the Aftis Fault, which exhibits right-lateral strike-slip motion and is oriented NW-SE (Figure 3.4c) (Abacha et al., 2023b).

7. Mila Sequence (July-August 2020) The Mila seismic sequence from July to August 2020 was characterized by the occurrence of two significant main shocks. The first event took place on July 17, 2020, at 08:12 with a magnitude of $M_w = 4.8$, located 1 km north of Sidi Merouane, near Grarem-Gouga. The second main shock occurred on August 7, 2020, at 06:15, registering a magnitude of $M_w = 5.0$ and detected 1 km north of Hamala, adjacent to the Beni-Haroune dam. This second event triggered a massive landslide that resulted in the destruction of approximately 100 illegally constructed houses situated on fragile terrain susceptible to landslides. Preliminary results from the HypoDD relocation of 981 events associated with the Mila sequence indicate a horizontal distribution characterized by two distinct swarms (Figure 3.5d, Boulahia, 2022).

8. Bejaia Sequence (2021) The latest sequence occurred off the coast of the Gulf of Béjaia following a strong earthquake with a magnitude of M_w 6.0. It included 420 events, including the main shock. Seismotectonic analysis revealed the presence of a thrust fault oriented approximately E-W, located at the western boundary of the WS segment (Figure 3.4d) (Yelles-Chaouche et al.,

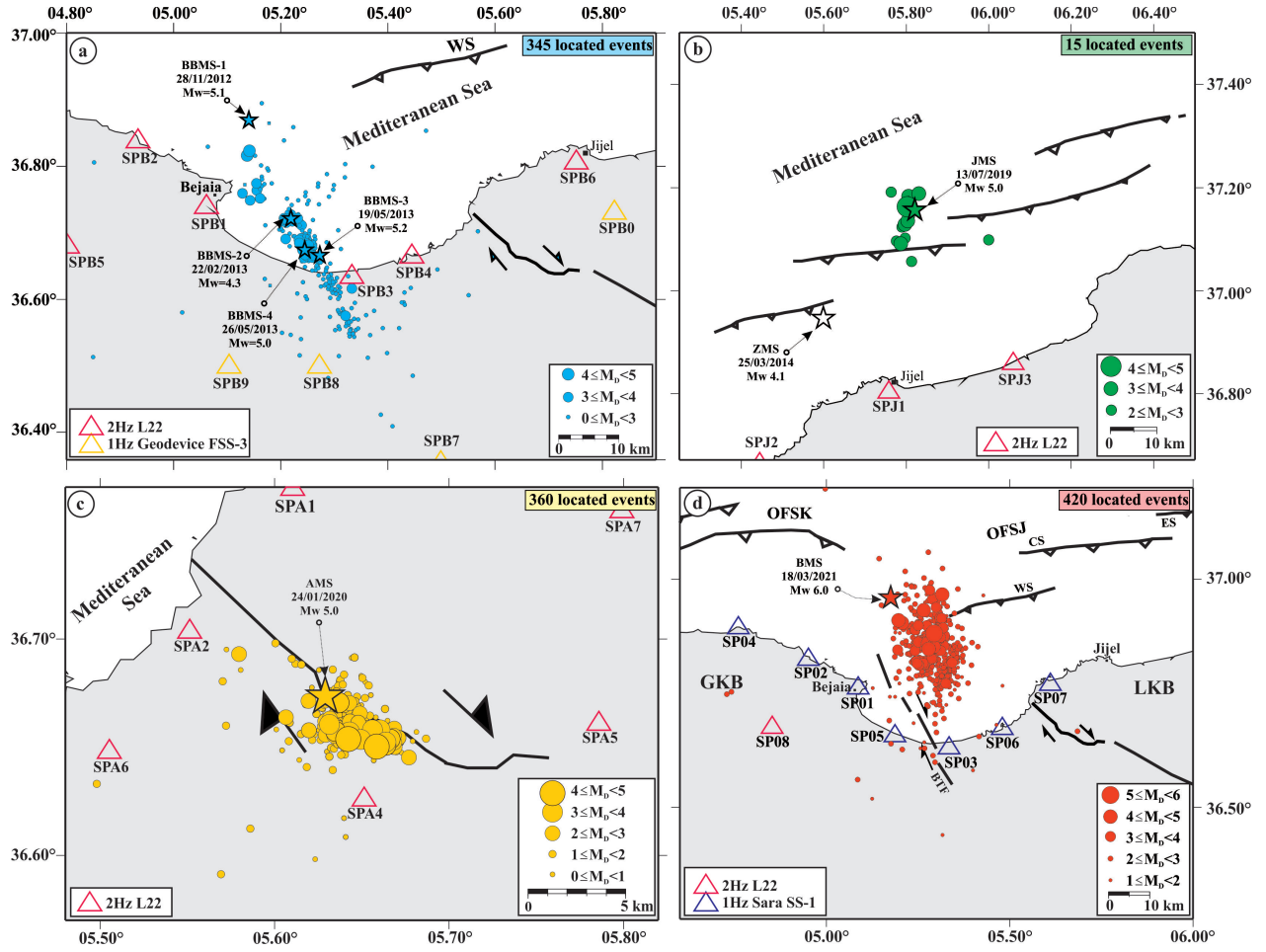


Figure 3.4: The horizontal distribution of the localized seismic events: **(a)** The Béjaia-Babors seismic sequence (2012-2013) [Boulahia et al., 2021](#). **(b)** The distribution of aftershocks from the 2019 Jijel earthquake [Yelles-Chaouche et al., 2021](#). **(c)** The El Aouana earthquake sequence in 2020 [Abacha et al., 2023b](#). **(d)** The Béjaia earthquake sequence in 2021 [Yelles-Chaouche et al., 2023](#).

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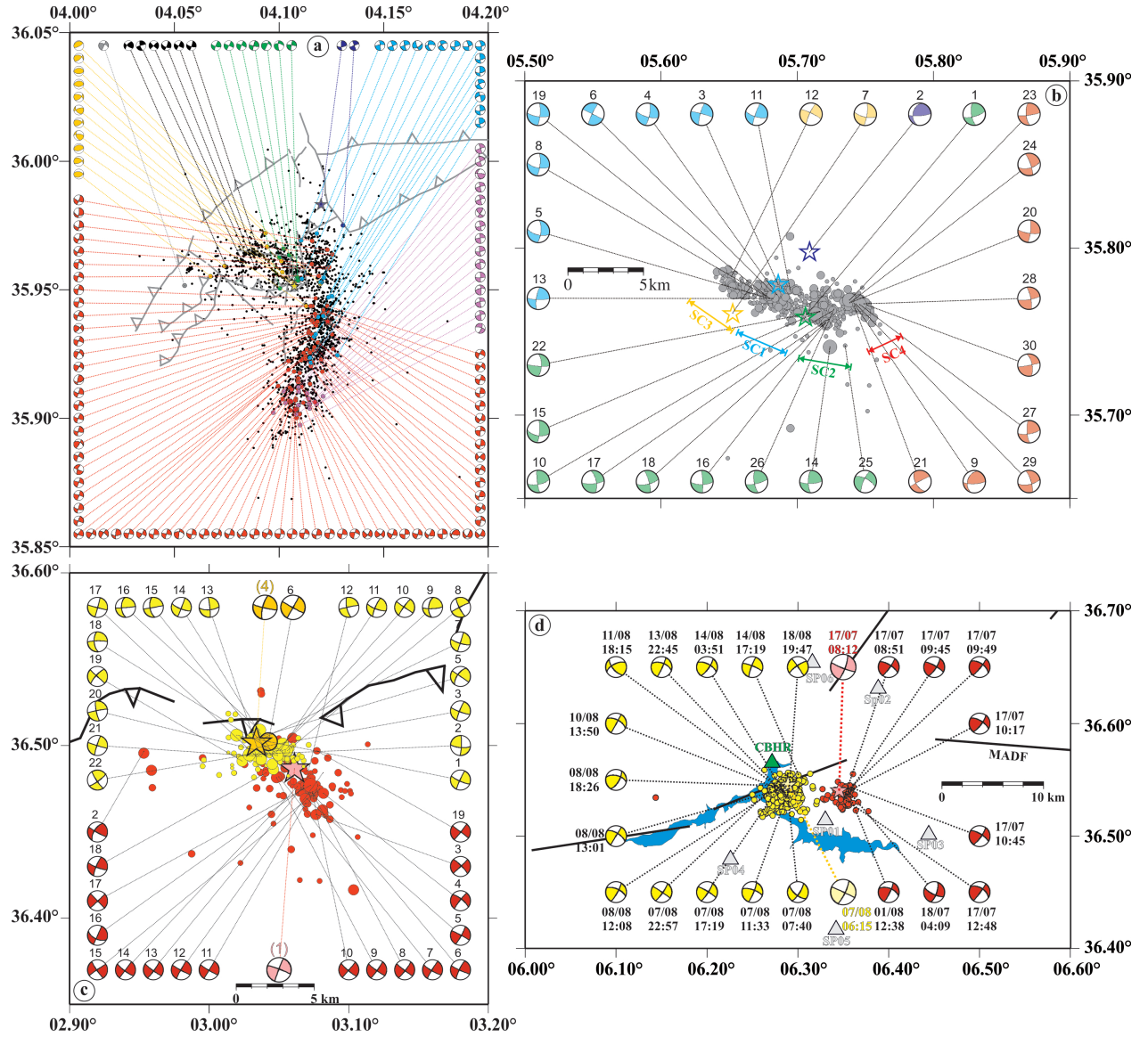


Figure 3.5: The horizontal distribution of the localized seismic events: (a) Beni Ilmane seismic sequence 2010 (Tikhamarine et al., 2024). (b) Ain Azzel seismic sequence 2015. (c) Hamam Melouane seismic sequence 2013 and 2014. (d) Mila seismic sequence 2020 (Boulahia, 2022).

3.1.3 Phases Identification

Seismic stations continuously record waveforms generated by earthquakes. To accurately determine the location and magnitude of an earthquake, it is essential to well identify the seismic phases and precisely record their arrival times. Once an earthquake's phases are identified at a station, the same methodology can be extended to multiple stations over longer time intervals. This broader dataset aids in identifying fault patterns and understanding seismic activity in a region.

The present study focuses on local earthquakes, characterized by prominent P and S wave phases.

1. **P Waves (Primary Waves):** - These are longitudinal waves that propagate through the Earth's medium via successive compressions and dilations (Figure 3.6). They are the fastest seismic waves and hence are the first to arrive at a seismic station and are easier to identify on the vertical component of the seismogram due to their polarization.. Due to their speed and lower amplitude, they are best recorded in low-noise environments, where their arrival times can be accurately noted. High levels of background noise can obscure the detection of P waves because their low amplitudes make them susceptible to interference.

2. **S Waves (Secondary Waves):** These transverse waves oscillate perpendicular to the direction of propagation(Figure 3.6). S waves travel slower than P waves but carry higher energy, often appearing as sharp, distinct peaks on a seismogram. Unlike P waves, S waves remain discernible even in noisier environments due to their higher amplitudes, although the exact onset time might be harder to pinpoint. Filters can enhance seismic wave identification under noisy conditions, ensuring better results for both phase identification and amplitude measurement.

3. **Surface Waves:** These waves travel along the Earth's surface and are generally slower than both P and S waves. There are two main types of surface waves: Love waves and Rayleigh waves. Love waves cause horizontal shearing of the ground, while Rayleigh waves produce a rolling motion, combining both vertical and horizontal ground movement. Surface waves usually have the largest amplitudes and can cause significant damage during earthquakes. Because of their slow speed and large amplitudes, they are often the most easily detected waves on seismograms, especially during large events.

3. **Coda Waves:** - After the body (P and S) and surface waves, seismic energy persists in the form of scattered waves known as coda waves. These waves result from scattering due to hetero-

geneities in the Earth's crust and are recorded during the later stages of the seismogram. Coda waves are crucial for estimating duration magnitudes. However, due to their weak signal strength and resemblance to background noise, coda waves often remain undetected unless processed carefully.

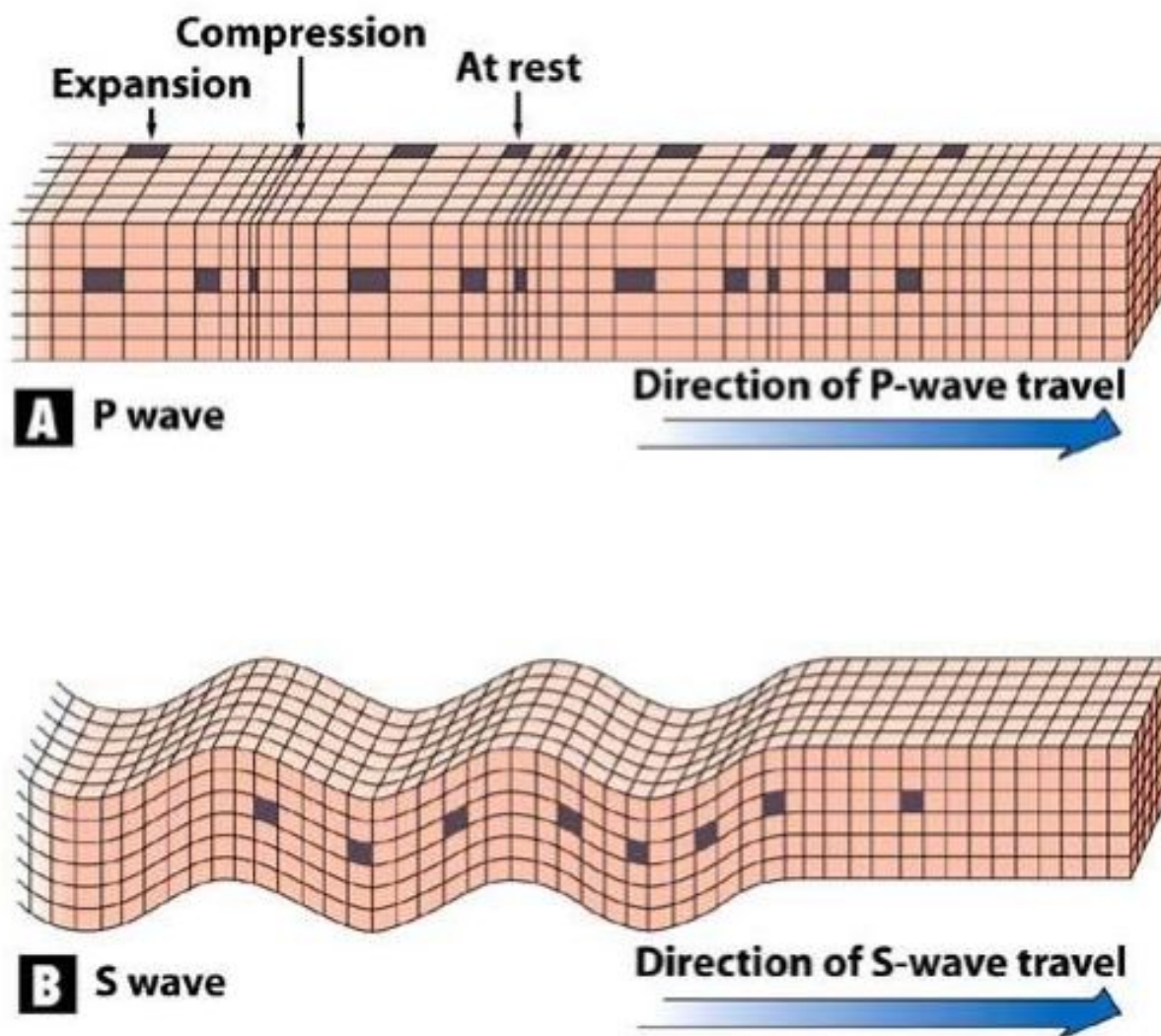


Figure 3.6: Particle motions associated with P and S waves (Shearer, 2009).

3.1.4 Development of a Continuous Seismic Database

In this sub-section, we describe the various processes used to create a continuous, complete, and structured database. This step is essential for processing seismic data. First, it's important to highlight that different digitizers paired with different sensors in temporary networks (e.g., GEOSTAR,

CUBE, GUARLP, etc...) generate data in various formats depending on the manufacturer: SEED, CUBE, GCF, DAT, etc. This diversity requires converting all formats to a standard, homogeneous format, SEED (Standard for the Exchange of Earthquake Data), which is maintained by IRIS (Incorporated Research Institutions for Seismology). In contrast, data from the permanent network are already recorded in SEED format.

The SEED format thus combines both the seismic time series collected by seismometers and the metadata files (dataless files) that store information related to the earthquake station siting, including the station geographic position, time series, instrument response, sampling frequency, and more. A dataless file can also provide a complete and comprehensive history of metadata for one or more network stations.

Seismic data are the digital records generated by the Analog-to-Digital (A/D) converter from the seismometer output. Keeping raw data and metadata separate provides security by allowing users the flexibility to apply the processing methods best suited to their study. The SEED format uniquely identifies data using four variables:

Network Code: A two-character identifier, e.g., "DZ" for the ADSN network.

Station Code: A four-character code, such as CKHR (Constantine Kef LaHmeR).

Location ID: A two-character identifier used when a station has been moved or repositioned within the ADSN.

Channel Code: A three-character code, where the first character represents the band type or sampling rate (E: Extremely Short Period, H: High Broad Band, B: Broad Band), the second indicates the instrument type (H: High Gain Seismometer, N: Accelerometer), and the third shows sensor orientation (N: North, E: East, Z: Vertical).

In practice, the SEED format was unsuitable for the data processing required for our analysis. Therefore, we converted all waveforms to the Seismic Analysis Code (SAC) format. Each SAC data file contains waveform data for a single component recorded at a single seismic station, along with a "header" that describes file contents, including data number of points, file type, sampling rate, and signal origin time. The raw data, converted to SAC format, are organized into annual directories (2010, ..., 2013, ..., 2024), with each annual directory containing daily folders (R001, R002, etc.). Each day is cataloged by hour (e.g., 2010.01.01-00.00.00, 2010.01.01-01.00.00), containing waveform data for a specified component from a given station.

Figure 3.7 summarizes the preprocessing steps needed to establish a clean, coherent, continuous waveform database.

Starting with our initial continuous database, we will create a new database focused on individual events. The first step is to generate a Trigger file, which contains each event's start time and signal duration. The file's name varies depending on the detection mode: for automatic detection, it is named with the extension (.hds), while for manual detection, it is named (.hdr). The next step involves extracting the events. A shell or bash script in Linux performs this task by reading each data from the Trigger file, and then cutting from the continuous data to isolate the signal for each event in SAC format.

The data are organized alphabetically by station, with SAC files saved in annual directories (2010, ..., 2024), followed by daily directories (R001, R002, etc.), and finally event-specific directories (e.g., 2010.06.19-15.58.22, 2010.07.21-16.01.43). The database created for this study includes over 8,000 earthquakes with magnitudes ranging from 1 to 6. Approximately 75% of these records are small events with magnitudes below 3.5.

3.1.5 Instrument Response Deconvolution

Deconvolution of the instrument response is a fundamental preprocessing step in seismology. It is applied to raw data, where the amplitude is initially uncalibrated in physical units. This process removes the influence of the seismometer from the raw recordings, restoring the ground motion the primary observable of interest. For recordings from a velocimeter, deconvolution produces a signal representing ground velocity, typically expressed in nm/s (following SAC conventions). By integrating this velocity signal, we can obtain the ground displacement, while differentiating it yields ground acceleration. This calibration is essential for accurate analysis and comparison across different seismic events and stations.

The instrument response represents the entire signal acquisition chain. In a sensor, the transducer (a magnet-coil pair) measures the position of a mass relative to the ground, converting displacement into an electrical voltage by induction. This voltage is proportional to ground velocity. The analog signal, recorded in volts, is first amplified and filtered with an anti-aliasing analog filter. It is then converted to a digital signal, after which it undergoes further filtering and resampling (see Figure 3.8).

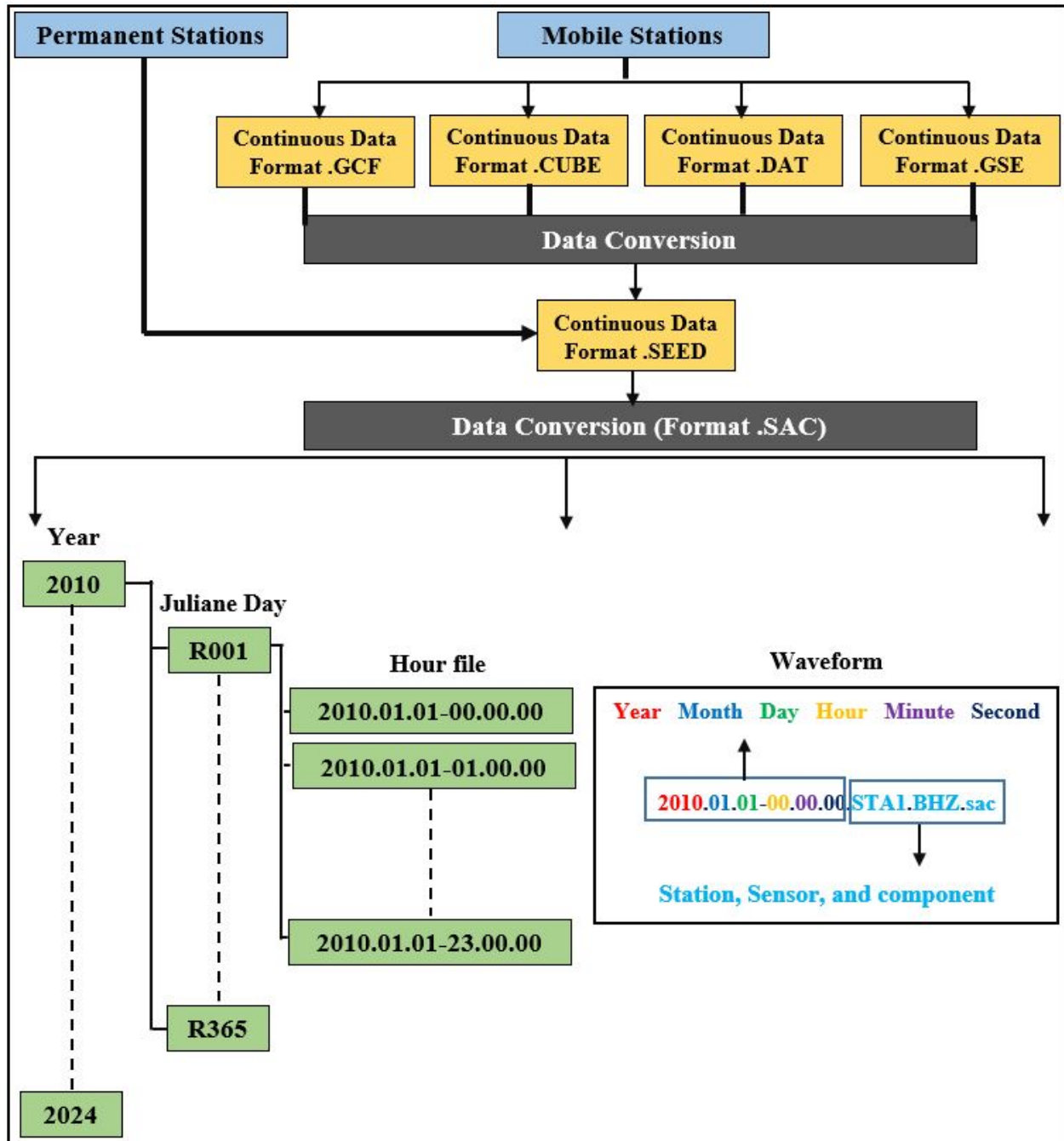


Figure 3.7: Steps and procedures for continuous database processing and data storage architecture

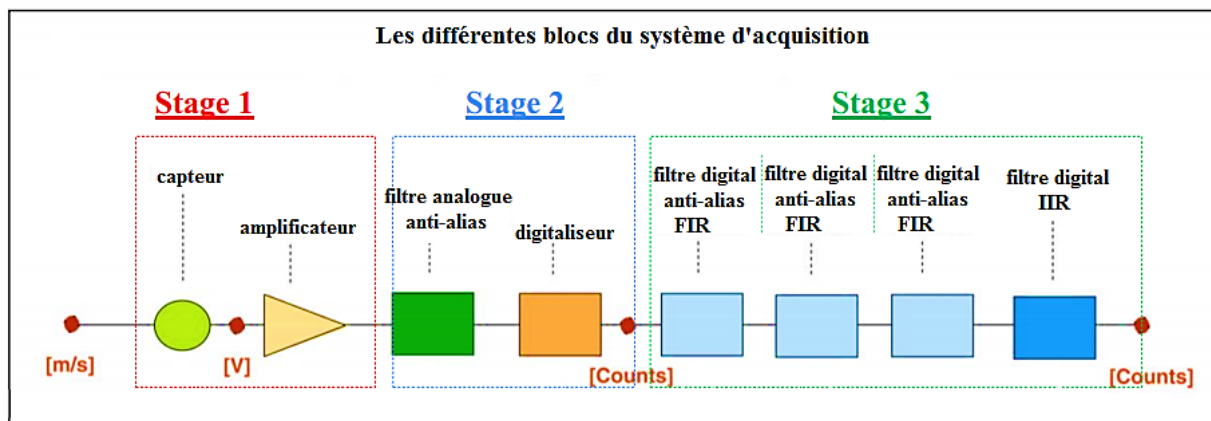


Figure 3.8: The various components of the acquisition system include conversion, amplification, and filtering, applied to ground motion (ground velocity in m/s) recorded by a seismological station (From Boulahia, 2022).

Each step in this process is documented in metadata files, which preserve the station's operational history. Known as RESP files, these metadata files are essential for instrument response deconvolution. All seismograms used in this work were deconvolved using the programs EvalResp and Polezeros. Although the sensors used are velocimeters, with signals reflecting ground velocity, the pole-zero files are expressed in terms of ground displacement. An example of a response file is shown in Figure 3.9.

By convention, RESP files include poles and zeros, gain, and the filtering-decimation stages associated with digitization. These files correspond to velocity measurements, as SAC also works by convention with velocities expressed in nm/s. Working directly in nm avoids the need for multiplying or dividing by large values. The sequence of SAC commands used for correcting the instrument response effect was automated with a Matlab program and shell scripts

In SAC, we obtain displacements using the command "transfer to none", velocities with "transfer to vel", and even accelerations with "transfer to acc". When using the pole-zero method, the command "from evalresp" can simply be replaced with "from polezeros".

The "freq" parameter in the "trans" command specifies a band-pass filter with corner frequencies $F1$, $F2$, $F3$, and $F4$ with $F1 < F2 < F3 < F4$, and $F4$ below Nyquist frequency. The taper is zero below $F1$ and above $F4$, and reaches unity between $F2$ and $F3$. $F1$ and $F2$ define the high-pass filter at low frequencies, while $F3$ and $F4$ specify the low-pass filter at high frequencies. Using the same pole-zero file with one additional zero applies integration (e.g., converting velocity to dis-

```

#          << IRIS SEED Reader, Release 5.2 >>
#
#          ===== CHANNEL RESPONSE DATA =====
B050F03    Station:      EMHD
B050F16    Network:     DZ
B052F03    Location:    00
B052F04    Channel:     BHZ
B052F22    Start date:  2006,335,00:00:00.0000
B052F23    End date:    No Ending Time
#
#          =====
#          +-----+-----+-----+-----+
#          +-----+-----+-----+-----+
#          +-----+-----+-----+-----+
#          +-----+-----+-----+-----+
#
B053F03    Transfer function type:      A [Laplace Transform (Rad/sec)]
B053F04    Stage sequence number:      1
B053F05    Response in units lookup:    M/S - velocity in meters per second
B053F06    Response out units lookup:   V - emf in volts
B053F07    A0 normalization factor:    63168.8
B053F08    Normalization frequency:    0.5
B053F09    Number of zeroes:           2
B053F14    Number of poles:            4
#
#          Complex zeroes:
#          i real      imag      real_error  imag_error
B053F10-13  0  0.000000E+00  0.000000E+00  0.000000E+00  0.000000E+00
B053F10-13  1  0.000000E+00  0.000000E+00  0.000000E+00  0.000000E+00
#
#          Complex poles:
#          i real      imag      real_error  imag_error
B053F15-18  0 -7.405000E-02  7.405000E-02  0.000000E+00  0.000000E+00
B053F15-18  1 -7.405000E-02 -7.405000E-02  0.000000E+00  0.000000E+00
B053F15-18  2 -1.777200E+02  1.777200E+02  0.000000E+00  0.000000E+00
B053F15-18  3 -1.777200E+02 -1.777200E+02  0.000000E+00  0.000000E+00
#
#          +-----+-----+-----+-----+
#          +-----+-----+-----+-----+
#          +-----+-----+-----+-----+
#          +-----+-----+-----+-----+
#
B058F03    Stage sequence number:      1
B058F04    Gain:                      9.999390E+02
B058F05    Frequency of gain:          5.000000E-01 HZ
B058F06    Number of calibrations:     0
#
#          +-----+-----+-----+-----+
#          +-----+-----+-----+-----+
#          +-----+-----+-----+-----+
#          +-----+-----+-----+-----+
#
B054F03    Transfer function type:      D
B054F04    Stage sequence number:      2
B054F05    Response in units lookup:    V - emf in volts
B054F06    Response out units lookup:   COUNTS - digital counts
B054F07    Number of numerators:       0

```

Figure 3.9: Example of the response file for the seismic station EMHD. The file includes information about the instrument response and calibration parameters.

placement), while removing a zero applies differentiation (e.g., converting velocity to acceleration). Temporal integration or differentiation of the signal produces consistent results.

3.2 Methodology

3.2.1 Local Magnitude (M_L)

In this subsection, we provide a detailed explanation of the methodology used for estimating the Local Magnitude (M_L). Unlike the original Richter method, which utilized only broadband stations, we employed both short-period and broadband stations to enhance the resolution and accuracy of our magnitude estimates. Instead of using the horizontal component of ground motion, we focused on the vertical component to better capture specific seismic wave characteristics. The instrumental response was also carefully removed during the data processing to ensure that the measured amplitudes accurately reflect the ground motion without the influence of the recording instruments. We outline the steps involved in this process, including data selection, waveform processing, and the specific formulae and parameters applied.

a- Use of Short-Period and Broadband Stations

The Algerian Digital Seismic Network (ADSN) consists of 72 permanent stations, including 20 broadband and 52 short-period stations. Especially for events below magnitude 2.8, the estimation of the M_L using only broadband stations is challenging, as highlighted by the study of [Boulahia, 2022](#). The earthquake detection capability of our region (see Figure 3.10) with a minimum of four broadband seismic stations is limited to $M=2.8$, for which moment magnitude (M_w) proves to be accurate. For smaller events, however, local magnitude (M_L) provides greater sensitivity and accuracy. Therefore, the integration of short-period stations enhances earthquake detection capabilities, is essential for improving seismic monitoring and accurately assessing seismic activity. To thoroughly evaluate the capability of short-period instruments in accurately measuring seismic amplitudes, particularly for local magnitude (M_L) calculations, we conducted an in-depth frequency response analysis of both broadband and short-period stations using Bode diagrams (see Figure 3.11). This analysis allowed us to assess the sensitivity of each instrument type across a range of frequencies, with a focus on their ability to capture waveform amplitudes. The results show

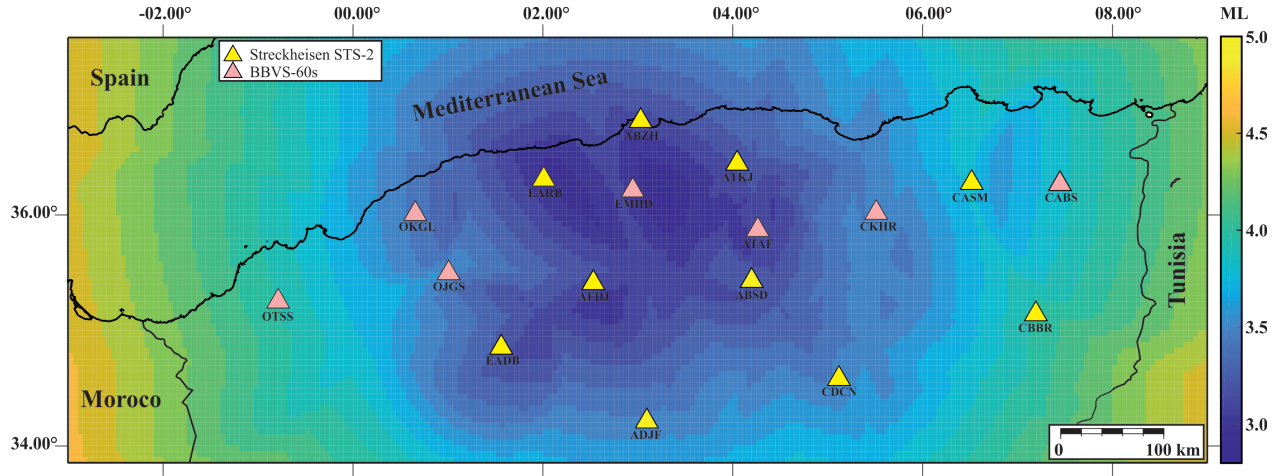


Figure 3.10: The capacity of earthquake detection using only Broadband stations with a minimum of four stations and Signal-to-Noise Ratio (SNR) > 3.

that both broadband and short-period stations exhibit effective sensitivity in the 1-20 Hz frequency range, which is critical for local magnitude estimation. Short-period instruments, in particular, demonstrated strong sensitivity to the high-frequency components of seismic waves, which are essential for accurately estimating maximum amplitudes a key factor in calculating M_L . In addition,

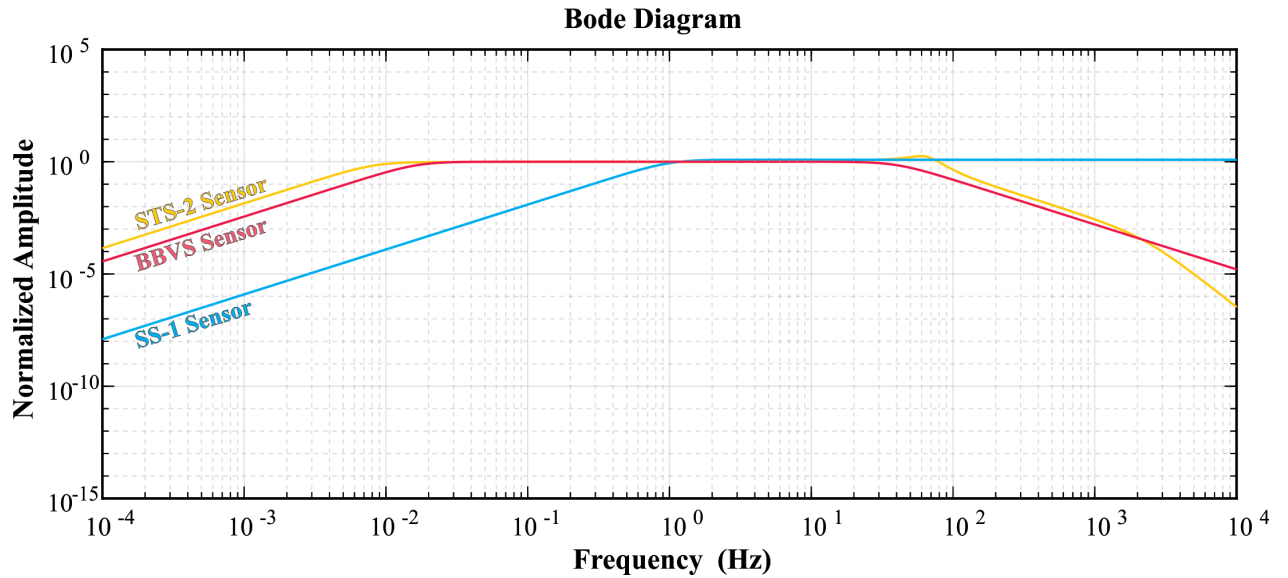


Figure 3.11: Bode Diagram: The normalized transfer functions of sensors SS-1, STS2, and BBVS 60s.

we performed a comparative analysis of co-located broadband and short-period instruments (see Figure 3.12) to directly evaluate the ability of short-period stations to estimate maximum amplitudes for local magnitude calculation. This comparison involved analyzing recordings from both

instrument types across a series of local seismic events. The results revealed a strong correlation between the maximum amplitudes recorded by broadband and short-period stations, confirming that short-period instruments are capable of accurately capturing the peak amplitudes necessary for M_L calculations. This consistency reinforces the reliability of short-period stations in local magnitude estimation and supports their use in seismic networks focused on monitoring local seismicity with high precision.

b- Use of the Vertical Component

The CRAAG uses maximum peak amplitude read on the vertical component in its M_L calculations for several reasons. Strategically placed mono-component (vertical) short-period sensors within the ADSN improve data availability, especially in regions with sparse station coverage. Furthermore, the vertical component often has lower background noise levels than the horizontal components, improving data reliability. Historically, for economic reasons only, since Algeria is a large country This approach is consistent with the practices of other researchers, including (e.g., [Ortega and Quintanar, 2005](#); [Bajc et al., 2013](#); [Condori et al., 2017](#); [Mahani and Kao, 2020](#)), who have successfully used the vertical component in M_L estimation. In addition, a comparison of the maximum vertical and horizontal WA amplitudes is presented in Figure 3.13 for a dataset of 3328 seismic records obtained from broadband stations only; there is minimal differences between the maximum amplitudes recorded on the vertical and horizontal components.

c- Inversion process

Fundamentally, the M_L scale establishes a simple empirical relationship for measured ground displacement. Data clouds to be averaged, as explained by [Boore, 1989](#), thereby facilitating M_L calculation. Implicit in the M_L equation presented by Equation 2.1 is the assumption that the scatter within the data cloud is random. Nevertheless, this assumption is not always valid since repeatable site effects can affect M_L . [Richter, 1958](#) realized this limitation and subsequently revised Equation 2.1 to include station corrections (S):

$$M_L = \log_{10}(A) - \log_{10}(A_0) + S \quad (3.1)$$

The station correction term (S) aims to correct for systematic and repeatable biases introduced on an instrument basis, depending on the ground conditions and instrument installation at each site,

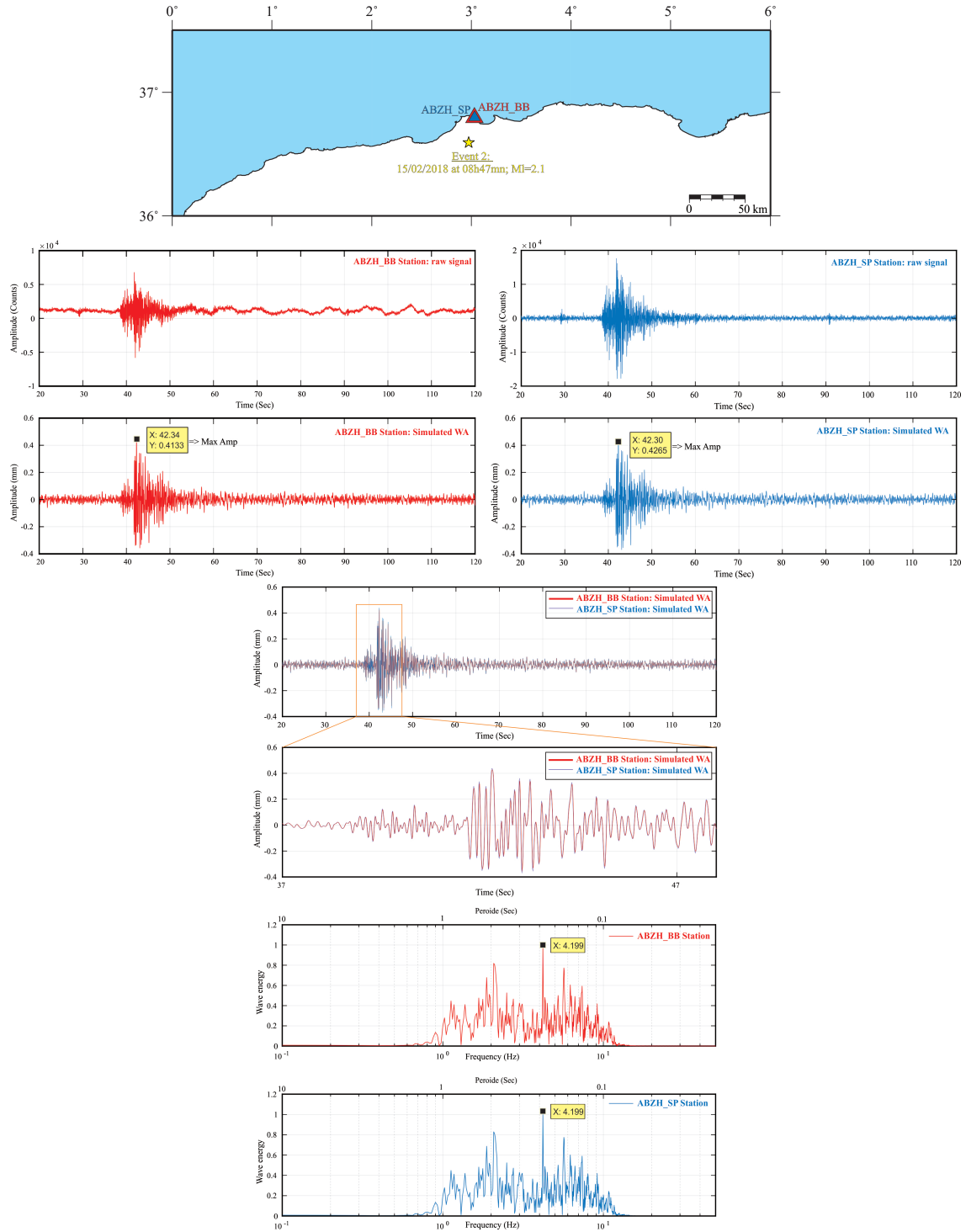


Figure 3.12: A comparison example between signals from two co-located seismic stations, one equipped with broadband BB (red one) and the other with short-period SP (blue one) seismological instruments, both situated at the same location, capturing data from the same local seismic event (April 1, 2018, M_L 3.7). The two signals are very similar even in terms of energy (the two last panels) on which we determine the dominant frequency.

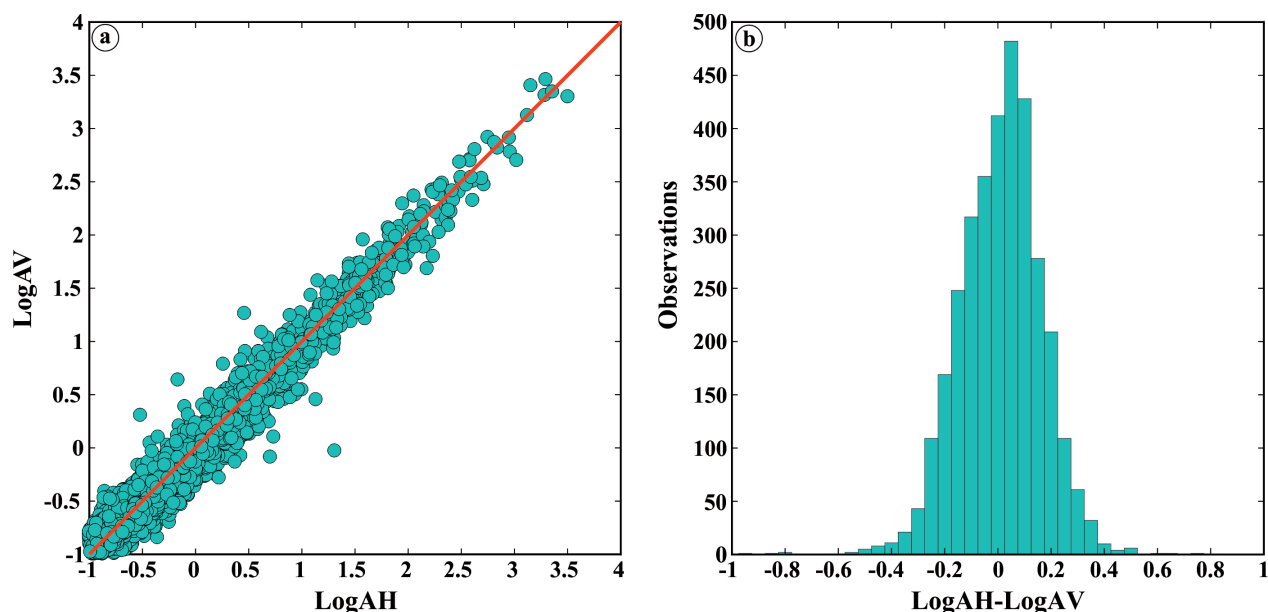


Figure 3.13: (a): Comparison of maximum amplitudes in millimeters for vertical (AV) and horizontal (AH) seismic data from broadband stations utilized in this study. The solid red line represents 1:1 line. (b): A histogram of the log differences between AH and AV

as outlined by [Richter \[1958\]](#). The attenuation correction can be determined, as well as the station correction term (S) and M_L values of the events, by solving a linear system of equations in the form of $d = Gm$ (e.g., [Menke, 1984](#); [Savage and Anderson, 1995](#); [Bindi et al., 2005](#)). The original scaling was established based on a Wood-Anderson displacement peak of 1 mm at a distance of 100 km, equivalent to an M_L of 3. Nevertheless, [Hutton and Boore \[1987\]](#) reviewed this method and advised adjusting local magnitude scales to the region being studied, they proposed fixing the scale at $M_L=2$, at a distance of 17 km, to attain greater flexibility when matching regional attenuation at longer distances. The recommendation was made due to concerns that [Richter, 1958](#) had mistakenly assumed a $1/R^2$ geometric spreading for distances < 30 km and that the average focal depth was excessively deep. Later on, [Alsaker et al., 1991](#) noted that adopting the reference point from [Hutton and Boore, 1987](#) caused a noticeable discrepancy in the attenuation correction for the Norwegian area. Careful consideration should be given when selecting a reference point, with emphasis on accurately reflecting the observed attenuation in the target region and across an appropriate distance range. In agreement with ([Richter, 1958](#); [Hutton and Boore, 1987](#)), [Alsaker et al. \[1991\]](#), anchored their scale at 60 km, where they had adequate data to make a sound judgment. It should be noted that despite these recommendations, only a few networks have adopted Hutton

and Boore suggestion. This hesitancy is likely due to the lack of sufficient near-field data at 17 km for most networks at the time of the study, making it difficult to justify its widespread adoption. Predominantly, two primary methodologies have been used for M_L calibration. The first, and most widely used, is the 'parametric model' introduced by [Bakun and Joyner, 1984](#). The second is the 'non-parametric model' as proposed by [Savage and Anderson, 1995](#).

- Parametric model

The parametric model provides an objective explanation of the term " $-\log_{10}(A_0)$ " by considering different factors, such as geometrical spreading, anelastic attenuation, and scattering, that cause a decrease in wave amplitude as seismic waves propagate. The equation to be minimized, as described in Equation 3.2, takes into account these effects.

$$-\log_{10}(A_0) = n * \log_{10}(R/R_{ref}) + b * (R - R_{ref}) + K_{ref} \quad (3.2)$$

In this context, "n" and "b" signify the geometrical spreading and anelastic attenuation terms, respectively. Collectively, these two coefficients collaboratively determine the attenuation correction (distance correction term). Moreover, "Rref" and "Kref" represent the reference distance and constant related to the M_L definition, respectively. The hypocentral distance (R) plays a fundamental role in the equation and is expressed in kilometers. Given the clearly defined and smooth functional form of the model presented in Equation 3.2, its determination requires only a small number of measurements, balancing the trade-off between accuracy and model complexity ([Langston et al., 1998](#)). Introducing more complexity to the parametric model is possible. The parameter $n * \log_{10}(R/R_{ref})$ can be replaced with a piecewise linear model, which can be implemented as either a bilinear or trilinear model based on the data characteristics. Furthermore, [Bindi et al., 2019](#) added complexity to the $n * \log_{10}(R/R_{ref})$ term by using a trilinear model and including two breakpoint distances. This modification provides more flexibility, especially for larger distances, as the piecewise model captures a more detailed regional attenuation than Equation 2.1 constraints. From Equations 3.1 and 3.2, the M_L relationship, defined as a piece-wise linear function, is characterized by two breakpoints at R_1 and R_2 as follows:

$$M_L = \begin{cases} \log_{10}(A_{ij}) + n_1 * \log_{10}(R_{ij}) + b * R_{ij} + S_j + C & \text{if } R_{ij} \leq R_1 \\ \log_{10}(A_{ij}) + n_1 * \log_{10}(R_1) + n_2 * \log_{10}(R_{ij}/R_1) + b * R_{ij} + S_j + C & \text{if } R_1 < R_{ij} \leq R_2 \\ \log_{10}(A_{ij}) + n_1 * \log_{10}(R_1) + n_2 * \log_{10}(R_2/R_1) + n_3 * \log_{10}(R_{ij}/R_2) + b * R_{ij} + S_j + C & \text{if } R_{ij} > R_2 \end{cases} \quad (3.3)$$

Where the constant $C = K_{ref} - b * R_{ref} - n_1 * \log_{10}(R_{ref})$.

The coefficients n_1 , n_2 , and n_3 represent the geometric spreading rates within three distance intervals defined by the transition distances between R_1 and R_2 . The indices "i" and "j" refer to the particular earthquake (i) and station (j). One possible way to approach Equation 3.3 is by formulating the system of linear equations in a standard matrix format and then solving the linear inversion problem ($\mathbf{d} = \mathbf{G} \cdot \mathbf{m}$). Here, the variable "d" corresponds to the known terms (data), specifically the $\log(A_{ij}) + K_{ref}$. The kernel matrix "G" establishes the relationship between "d" and "m", where "m" encompasses the unknown parameters (the model) such as n_1 , n_2 , n_3 , b , M_{Lij} , and S_j . Importantly, all these unknown parameters are computed simultaneously. To address the unknown site effects, station correction factors are relative to the site conditions. This involves constraining the S terms to zero when summed across all stations. This strategic constraint significantly diminishes discrepancies in the M_L model across various stations, contributing to a more robust and consistent analysis. when we have a total of "p" earthquakes and "q" stations, the universal structure of the

kernel matrix "G", along with the vectors "d" and "m", can be expressed as follows:

$$\begin{pmatrix} \log_{10}(A_{11})+3 \\ \log_{10}(A_{12})+3 \\ \log_{10}(A_{13})+3 \\ \vdots \\ \log_{10}(A_{21})+3 \\ \log_{10}(A_{22})+3 \\ \log_{10}(A_{23})+3 \\ \vdots \\ \log_{10}(A_{pq})+3 \\ 0 \end{pmatrix} = \begin{pmatrix} -\log_{10}\left(\frac{R_{11}}{100}\right) & 0 & 0 & -(R_{11}-100) & 1 & 0 & \dots & 0 & -1 & 0 & 0 & \dots & 0 \\ -\log_{10}\left(\frac{R_1}{100}\right) & -\log_{10}\left(\frac{R_{12}}{R_1}\right) & 0 & -(R_{12}-100) & 1 & 0 & \dots & 0 & 0 & -1 & 0 & \dots & 0 \\ -\log_{10}\left(\frac{R_1}{100}\right) & -\log_{10}\left(\frac{R_2}{R_1}\right) & -\log_{10}\left(\frac{R_{13}}{R_2}\right) & -(R_{13}-100) & 1 & 0 & \dots & 0 & 0 & 0 & -1 & \dots & 0 \\ \vdots & \vdots & \vdots & \vdots & \vdots & \vdots & \vdots & \vdots & \vdots & \vdots & \vdots & \vdots & \vdots \\ -\log_{10}\left(\frac{R_{21}}{100}\right) & 0 & 0 & -(R_{21}-100) & 0 & 1 & \dots & 0 & -1 & 0 & 0 & \dots & 0 \\ -\log_{10}\left(\frac{R_1}{100}\right) & -\log_{10}\left(\frac{R_{22}}{R_1}\right) & 0 & -(R_{22}-100) & 0 & 1 & \dots & 0 & 0 & -1 & 0 & \dots & 0 \\ -\log_{10}\left(\frac{R_1}{100}\right) & -\log_{10}\left(\frac{R_2}{R_1}\right) & -\log_{10}\left(\frac{R_{23}}{R_2}\right) & -(R_{23}-100) & 0 & 1 & \dots & 0 & 0 & 0 & -1 & \dots & 0 \\ \vdots & \vdots & \vdots & \vdots & \vdots & \vdots & \vdots & \vdots & \vdots & \vdots & \vdots & \vdots & \vdots \\ -\log_{10}\left(\frac{R_1}{100}\right) & -\log_{10}\left(\frac{R_2}{R_1}\right) & -\log_{10}\left(\frac{R_{pq}}{R_2}\right) & -(R_{pq}-100) & 0 & 0 & \dots & 1 & 0 & 0 & 0 & \dots & -1 \\ 0 & 0 & 0 & 0 & 0 & 0 & \dots & 0 & -1 & -1 & -1 & \dots & -1 \end{pmatrix} \begin{pmatrix} n_1 \\ n_2 \\ n_3 \\ b \\ M_{L1} \\ M_{L2} \\ \vdots \\ M_{Lp} \\ S_1 \\ S_2 \\ S_3 \\ \vdots \\ S_q \end{pmatrix} \quad (3.4)$$

Terms related to geometric spreading Terms related to anelastic attenuation Event term Station term

- Non-parametric model

The non-parametric method involves an empirical estimation of " $-\log(A_0)$ " through the analysis of recorded data. This estimate is resolved at each node by linear interpolation from the available data. The model is designed to flexibly adapt to complex attenuation scenarios and to account for significant variations in attenuation with distance, as demonstrated by [Lolli et al., 2015](#). Furthermore, the inclusion of constraints ensures that the inverted attenuation model retains a distinct identity separate from station corrections and magnitudes. Unlike the other methods that rely on predefined distance ranges, this approach involves inversion of the equation over a range of distances. While it does not provide a direct physical interpretation, it does provide a robust empirical framework for characterizing attenuation patterns. The nonparametric model as described by [Savage and Anderson, 1995](#) is summarized in Equation 3.5:

$$\log_{10}(A_{ij}(r_{ij})) = -a_n \cdot \log_{10} A_0(R_n) - a_{n+1} \cdot \log_{10} A_0(R_{n+1}) + b_i \cdot M_{Li} - C_j S_j \quad (3.5)$$

Where: $a_n = \frac{R_{n+1} - r_{ij}}{R_{n+1} - R_n}$

In this context, a_n represents a coefficient used in linear interpolation to node n at a distance denoted by R_n . The subsequent coefficient, a_{n+1} , is expressed as $1 - a_n$. Variables b_i and C_j act as dummy variables, taking values of 1 or 0 depending on the existence or non-existence of the event-station pair. " n " is the number of nodes that separate the distance bins, and the hypocentral distance r_{ij} is such that $R_n \leq r_{ij} < R_{n+1}$. M_{Li} is the local magnitude of earthquake " i ", S_j is the station correction of station " j ". The linear system is subsequently solved, adhering to the specified constraints: $-\log_{10}(A_{ref}) + M_{Lref} = -\log_{10} A_0(R_{ref})$ where A_{ref} is some reference WA displacement and $\sum S_j = 0$.

For a more detailed and thorough understanding of all the steps involved in the estimation of local magnitude, please refer to Chapter 4, which offers an in-depth explanation of the methodology, covering data collection, processing techniques, and the calculations required to derive local magnitude.

3.2.2 Duration magnitude (M_d)

[Aki and Chouet, 1975](#) proposed that coda waves resulting from local earthquakes are often inter-

puted as waves that have been back-scattered by a multitude of heterogeneities uniformly dispersed throughout the crust. Consequently, for a given local earthquake occurring at epicentral distances less than 100 km, the total duration of a seismogram appears to be nearly independent of both distance and azimuth, as well as of the structural complexities along the direct wave path from the seismic source to the recording station. Additionally, the coda envelope configuration, which is characterized by exponential decay over time, remains constant. The magnitude of the earthquake is the primary factor that affects the amplitude level of the coda envelope and the duration of the seismic signal.

a- Automatic Duration Estimate

This technique, specifically designed to measure the duration of earthquakes, relies on estimating the signal-to-noise ratio (SNR) across seismic records. The method is designed to assess the temporal extent of seismic events by analyzing the balance between the discernible signal and background noise within the recorded data. Event duration, denoted as τ (Equation 2.2), is determined by considering the vertical component of ground velocity records. This approach follows the initial framework outlined by [Real and Teng, 1973](#). Duration is defined as the time elapsed from the initial P-wave arrival to the point along the trace where the amplitude of the signal coda diminishes to the level of background noise. The duration is pinpointed when the SNR falls below the threshold of 1.05 within the moving window.

b- Envelope calculation

The Hilbert transform offers a useful approach to evenly sampled signals, allowing for the creation of the analytic signal, a complex representation that excludes negative frequencies. This complex 'analytic signal' simplifies analysis by separating amplitude from frequency information. The signal envelope, which contains amplitude details, outlines the overall signal shape. Meanwhile, the frequency component reveals the instantaneous frequency, which represents the mean frequency at a specific point in time. This separation of amplitude and frequency improves the clarity and interpretability of signal characteristics. Given a modulated waveform $f(t)$ and its Hilbert transform denoted as $H[f(t)]$. The computation of the analytic signal involves subtracting the Hilbert transform of the function multiplied by the imaginary unit " i ".

$$f_a(t) = f(t) - i.H[f(t)] \quad (3.6)$$

The analytic signal $f_a(t)$ is a complex helical sequence, where the real part represents the original data and the imaginary part embodies the Hilbert transform. The Hilbert transform is essentially the real signal with a 90-degree phase shift. Therefore, if the real part of the signal is a sine function, the imaginary part becomes a cosine function, and vice versa. The formulas below can be used to obtain the envelope, phase, and instantaneous frequency from the analytic signal:

$$Envelope = \sqrt{Real(f_a(t))^2 + Imag(f_a(t))^2}$$

$$Phase = \arctan\left(\frac{Imag(f_a(t))}{Real(f_a(t))}\right)$$

$$Inst.freq = \frac{\delta Phase}{\delta t}$$

The formula mentioned earlier is used to derive the envelope, which provides the overall shape of the signal by touching only at its maxima. The envelope encapsulates the combined amplitude of all frequencies present in the signal while disregarding specific frequency details. The upper envelope is obtained using the given formula, and the lower envelope is obtained by taking the negative of the same formula. This pair of upper and lower envelopes provides a complete representation of the amplitude dynamics of the signal, capturing its changes over time. The phase will continuously increase, except when sudden jumps in frequency content occur. Changes in slope indicate frequency shifts, providing a clearer indication of changes compared to instantaneous frequency. Instantaneous frequency tends to fluctuate rapidly due to minor phase adjustments. The instantaneous frequency represents the mean frequency of the signal at a specific time, measured in radians per time unit (to be adjusted for Hz). Its advantage is providing a singular value. In contrast, a spectrogram makes it easier to observe how frequency changes over time but is challenging to quantify and requires more computation time. Instantaneous frequency often requires smoothing for clarity. A small, sudden phase jump can cause a significant instantaneous frequency leap, even

if it comes from a relatively insignificant part of the signal, such as the noise before an earthquake.

For an overview of the process used in estimate duration magnitude, see Chapter 5, which details the procedures, including data analysis, time window selection, and formulas applied.

3.3 Conclusion

In conclusion, the processes outlined in this chapter are key to transforming raw seismic data into accurate, interpretable information about earthquake characteristics. By carefully identifying seismic phases and leveraging standardized data formats, we ensure that magnitude calculations and event locations are accurately determined. The creation of a continuous, structured seismic database facilitates the efficient processing of large datasets, while instrument response deconvolution ensures that ground motion is properly represented. These steps form the foundation for any advanced seismic study, providing a reliable framework for understanding seismic events and their impact. The automation of these processes further enhances efficiency, allowing for the analysis of thousands of events across multiple stations. Ultimately, the techniques presented here contribute to more precise earthquake characterization, enhancing our understanding of seismic activity in the studied regions.

Chapter 4

A Local Magnitude Scale (M_L) for Northern Algeria

4.1 Introduction

This section is derived from a manuscript published in the *Tectonophysics* journal, authored by **Roubeche Khaled, Fethi Semmane, Abacha Issam, Boulahia Oualid, Rahmani Sofiane, and Tikhamarine El Mehdi**. The study offers new insights into seismic activity in Northern Algeria, focusing on the development and validation of the region's first local magnitude scale (M_L), which allows for more precise earthquake magnitude assessments.

The magnitude is a fundamental seismological parameter that is closely linked to both earthquake shaking and seismic hazard. It is an integral part of numerous scientific activities, such as the creation of earthquake catalogs used, among other things, for the statistical analysis of earthquakes. These studies are essential for seismic hazard assessment and the calculation of parameters such as the Gutenberg-Richter b value ([Wiemer and Wyss, 2002](#); [Wyss et al., 2004](#); [Ghosh et al., 2008](#)). Since 1935 ([Richter, 1935](#)), a variety of magnitude scales have been developed. Richter magnitude (also known as Local Magnitude, M_L) continues to be widely used as a significant reference point due to its simplicity and global reach, particularly in the media. M_L quantifies energy released by local earthquakes and evaluates seismic activity in specific regions (e.g., [Richter, 1958](#); [Kanamori and Jennings, 1978](#); [Hutton and Boore, 1987](#); [Boore, 1989](#)). The local magnitude scale was originally created by measuring the highest horizontal amplitude, specifically from the zero to the peak

of ground displacement on a Wood-Anderson seismogram. The Wood-Anderson seismometer is a horizontal torsion design with a natural period of 0.8 seconds, a magnification of 2800, and a damping constant of 0.8. This design served as the cornerstone of Richter's M_L scale. However, its use outside California was initially limited due to the scarcity of standard Wood-Anderson torsion seismographs in other regions. The Wood-Anderson seismogram can be efficiently synthesized from a variety of broadband or short-period digital waveforms. Data from modern digital stations can be used to simulate the response of the Wood-Anderson seismograph by applying deconvolution techniques to their instrumental responses and convolution with the Wood-Anderson transfer function. In particular, certain regions have discovered that using variable damping factors and magnification values (e.g., damping constant 0.7 and magnification 2080) can lead to more accurate simulations of Wood-Anderson behavior (e.g., [Uhrhammer and Collins, 1990](#); [Uhrhammer et al., 1996](#)). While these differences are minor, they could result in a maximum deviation of 0.13 magnitude units while computing local magnitude. Richter's definition states that a magnitude 3 earthquake at a distance of 100 km would produce an amplitude of 1 millimeter on a Wood-Anderson seismogram, mainly for crustal earthquakes in southern California. For M_L calculations to accurately reflect the behavior of seismic attenuation in regions other than Southern California, it is essential to properly project the ground motions recorded at the reference distance. This process is known as magnitude calibration ([Bobbio et al., 2009](#)). Numerous global studies have been carried out to calibrate local magnitude scales, such as those in Italy ([Bobbio et al., 2009](#), [Di Bona, 2016](#)), in Turkey ([Bindi et al., 2007](#), [Kılıç et al., 2017](#)), in the United Kingdom ([Ottemöller and Sargeant, 2013](#)), in New Zealand ([Ristau et al., 2016](#)), in South Egypt ([Abdullah et al., 2022](#)), in Slovenia ([Bajc et al., 2013](#)), in Greece ([Scordilis et al., 2016](#)), and in the Ethiopian Plateau ([Brazier et al., 2008](#)). To date, Algeria has not yet established a local magnitude scale specific to its region. The Algerian Digital Seismic Network (ADSN) started in 2006 ([Yelles-Chaouche et al., 2013](#)), and since then, the calculation of M_L has been based on the relationship developed for southern California ([Hutton and Boore, 1987](#)), using the vertical components. However, this approach has led to an underestimation of earthquake magnitude (0.2 units) compared with other scales. It is therefore imperative to use a more accurate M_L that corresponds well to actual ground shaking levels for a better assessment of earthquake hazards in regions of rapid population growth, such as northern Algeria. For several years, the earthquake monitoring department at the Center for Research in As-

tronomy, Astrophysics, and Geophysics (CRAAG) and other articles by Algerian researchers (e.g., Semmane et al., 2012; Yelles-Chaouche et al., 2014), used an unpublished relationship inspired by Lee et al., 1972 for the computation of the duration magnitude. This type of magnitude has recently been adopted for the compilation of the new Algerian seismic catalog covering the years 2006 (since the start of the Algerian digital seismic network) to 2020, as documented in the work of Yelles-Chaouche et al., 2022. The study conducted by Bellaleme et al., 2022 employed empirical relationships to convert various intensity and magnitude scales (I_{max} , m_b , m_{bLg} , M_s , and M_L) provided by international agencies, like ISC, IGN, and NEIC, to moment magnitude (M_w). Furthermore, they proposed new empirical relationships based on the General Orthogonal Regression method (GOR) to convert M_L -CRAAG and M_d -NIMT (Tunisian national seismological agencies) into M_w . Unfortunately, establishing reliable empirical relationships currently faces significant difficulties due to data heterogeneity, bias, and systematic error. Differences in data collection methods, geological conditions, and instrumentation between local and international agencies can complicate the process of establishing a reliable empirical relationship. The existence of bias and systematic error in the values calculated by local networks poses a significant problem. If local magnitude values are questioned due to uncorrected biases or errors, their incorporation into an empirical relationship using international values may propagate these inaccuracies, leading to unreliable results. In light of these challenges, this study focuses on estimating the M_L scale from the amplitude inversion of 1901 ADSN-recorded earthquakes that occurred in northern Algeria between January 1, 2010 (when the current network geometry was completed), and June 1, 2022, with a magnitude range of 1–5.8, and hypocentral distances of 5–600 km. The results of this chapter will be used to improve earthquake magnitude calculations in the region. The second objective of this chapter is to examine the relationship between M_L and M_w using good-quality data recorded only by the Algerian Digital Seismic Network (ADSN).

4.2 Seismic Data

The waveforms were obtained from 72 permanent seismic stations within the Algerian network, resulting in a total of 17377 trace recordings extracted from 1901 earthquakes that occurred between January 1st, 2010, and June 1st, 2022. These seismic events were captured by a minimum

of five stations and spanned a geographic area ranging from -2° to 8.5° longitude and 33.5° to 37° latitude (Figure 4.1a). The magnitude of these earthquakes ranged from 1 to 5.8, with a maximum depth of 25 km and hypocentral distances from 5 to 600 km (Figure 4.1b, c, and d). For most events (mainly for the eastern part), hypocentral parameters, including longitude, latitude, depth, and hypocentral distance, were obtained from various studies (e.g., [Bendjama et al., 2021](#); [Boulahia et al., 2021](#); [Boulahia, 2022](#); [Abacha et al., 2023a](#); [Abacha et al., 2023b](#)), and for the remaining events we extracted hypocentral parameters from [Yelles-Chaouche et al., 2022](#). To ensure high-quality data selection, we excluded events occurring within 25 seconds of each other to prevent overlapping seismograms at some stations. Additionally, events with depths exceeding 25 km were excluded, as such depths are rare in northern Algeria and can often be attributable to artifacts as noted by [Bajc et al., 2013](#). Moreover, the data had to exhibit a signal-to-noise ratio of at least three. The SNR is determined by calculating the average ratio of the spectral amplitudes of the P-wave (window around P-wave) to the noise (window before P-wave) in the spectral domain (Figure 5.1). This process involves dividing the spectral amplitudes of the P-wave by those of the noise. The zero-to-peak amplitude was calculated from the vertical component recordings. The majority (72%) of the velocimeter seismic stations used in this study are short-period stations equipped with only one vertical component, while the remaining are broad-band stations with three components. Figure 4.1a illustrates the distribution of seismic stations, events, and the ray path pattern used for this study. The study region's higher observation density and the presence of high seismicity areas ensured that the ray path coverage was considered satisfactory. Richter's methodology only considered the horizontal components. However, research by ([Alsaker et al., 1991](#); [Bindi et al., 2005](#); [Saunders et al., 2013](#)) has shown minimal differences between the maximum amplitudes recorded on the vertical and horizontal components, particularly for short distance events. Furthermore, the use of the vertical component has been shown to provide a fairly accurate estimate of the M_L . In addition, the horizontal component amplification tends to be more pronounced for stations located on unconsolidated soils, while the vertical component is less affected. To evaluate this further, a comparison of the maximum vertical and horizontal displacement amplitudes is presented in Figure 3.13 for a dataset of 3328 seismic records obtained from broadband stations only. The mean value of the vertical-to-horizontal ratio is 0.96, with a high correlation coefficient of 0.97. To accurately determine the maximum amplitude, a data processing procedure was employed. Initially, a time

window spanning 20 seconds before and 250 seconds after the arrival of the P-wave was defined. Within this window, both the mean and linear trends were removed. Subsequently, the original recording's instrument response was deconvolved, and the resulting corrected waveform was convolved with a simulated Wood-Anderson (WA) instrument response using a damping factor of 0.7, a period of 0.8 seconds, and a gain factor of 2080, as described in [Uhrhammer and Collins, 1990](#); [Uhrhammer et al., 1996](#).

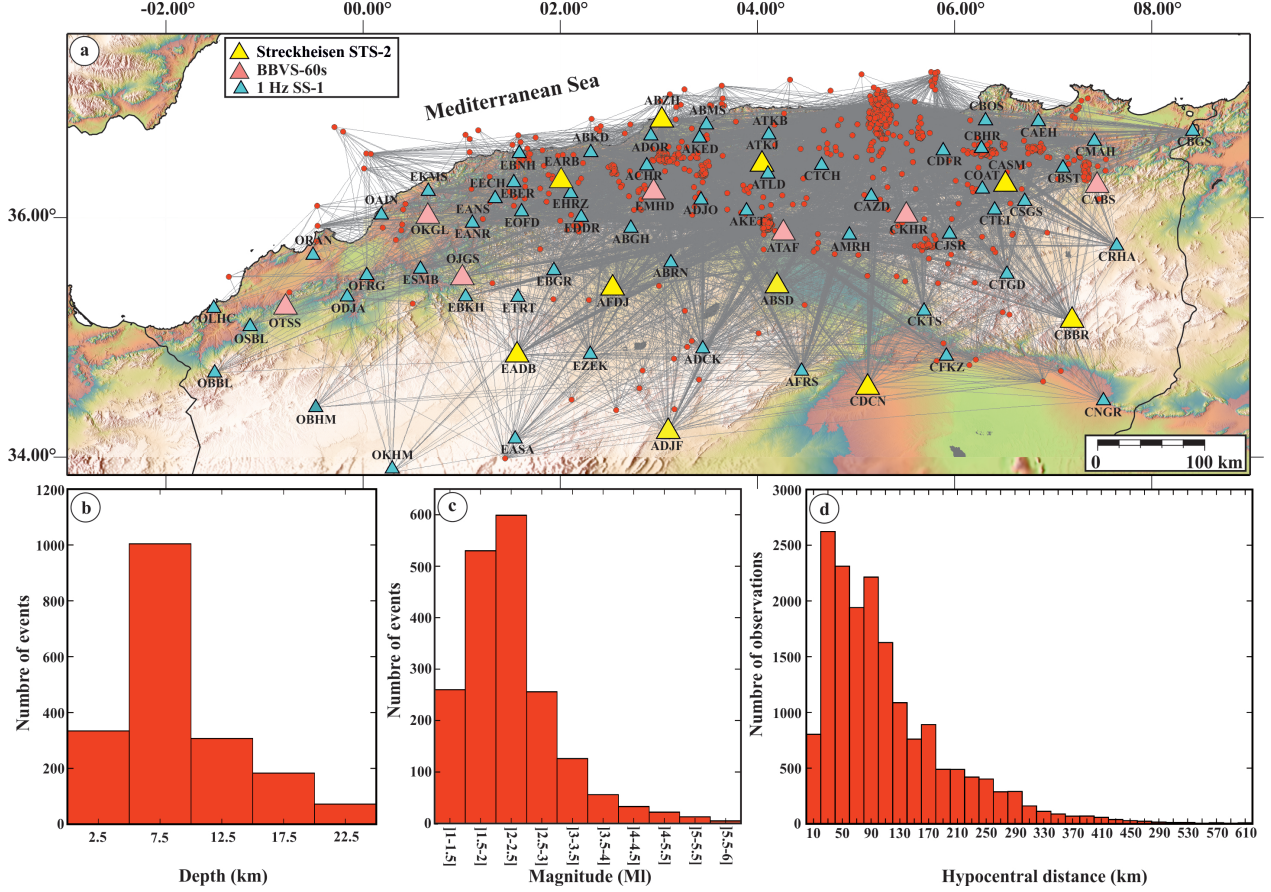


Figure 4.1: (a) Representation of the study area. Red circles denote earthquake locations. Triangles indicate station positions. Gray lines represent the ray paths; (b) Depth distribution of used earthquakes; (c) Distribution of magnitude levels; (d) Distribution of hypocentral distances.

4.3 Methodology

We aim to investigate regional attenuation effects in northern Algeria by analyzing amplitude decay with distance. To determine whether single, bilinear, or trilinear attenuation models are appropriate for our data, we employ a method based on ([Yenier, 2017](#)). To ensure that earthquakes of varying

magnitudes are measured consistently, we adjust their amplitudes to a common level using three distance ranges where seismic activity is well-recorded: 10-40 km, 20-80 km, and 100-150 km (Figure 4.1d). We achieve this by calculating the geometric mean of the amplitudes for each event that has at least four measurements in the reference distance ranges and its magnitude must exceed 2.5. All event amplitudes are then normalized accordingly. We repeat this process for each event and each reference distance interval. The average normalized amplitudes for all events should reach one at the reference distance intervals. This step eliminates source size variations and only reflects the area's attenuation characteristics. Note that site-effect are still included in the normalized amplitudes. The normalized amplitude results for the three distance intervals are shown in Figure 4.2.

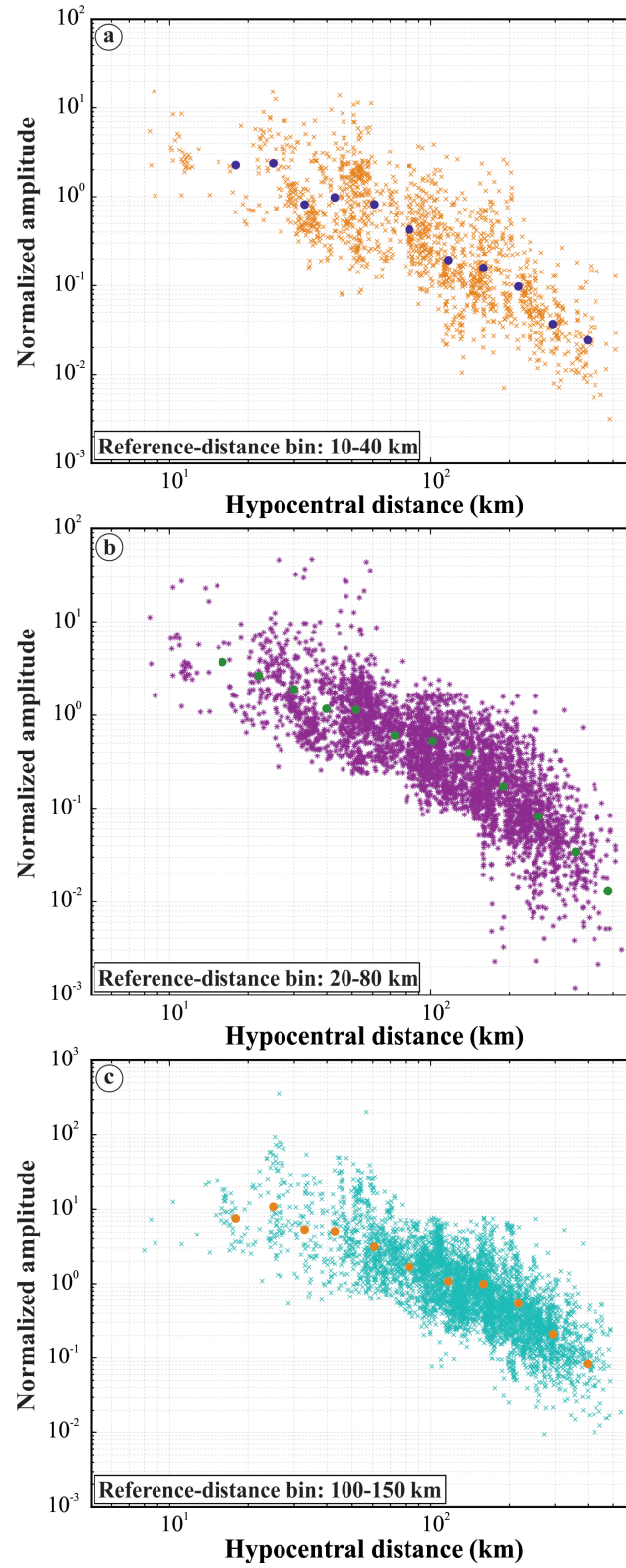


Figure 4.2: (a and b) The decay of normalized earthquake amplitudes with distance is illustrated in each panel, with the lower corner indicating the reference distance range for which at least three amplitude measurements were taken. The large circles in the graph depict the logarithmic average of normalized amplitude values calculated for log-spaced distance intervals.

The local magnitude, as defined by Richter in 1935 is given by equation 3.1 as (see section 3.2.1):

$$M_L = \log_{10}(A) - \log_{10}(A_0) + S \quad (4.1)$$

Where A represents the maximum trace amplitude in millimeters from a Wood-Anderson seismogram, (S) is the station correction factor, and $-\log_{10}(A_0)$ represents the distance correction term and can be defined using the equation proposed by [Bakun and Joyner, 1984](#) (Equation 3.2). Preliminary analysis of amplitude decay clearly shows that the observed amplitudes exhibit Moho-bounce behavior in the range of around 90 to 200 kilometers (Figure 4.2). As a result, we have chosen to use a trilinear model to account for distance correction terms during the data fitting process. Within the framework of this model, the general M_L expression by equation 3.3.

By definition, the condition specified in Equation 3.1 is that when a seismic event is recorded by a Wood-Anderson instrument located 100 km from the hypocenter and exhibits an amplitude of 1 mm, the corresponding local magnitude is 3.0. To fulfill this criterion, we have selected a reference distance R_{ref} of 100 km and set the constant K_{ref} equal to three. One approach to address Equation 3.3 involves formulating the system of linear equations in a standard matrix format and subsequently solving the linear inverse problem ([Menke, 1984](#)) ($d = Gm$). Recognizing the uncertainty of site effects, the inversion problem was constrained by considering the sum of station corrections to be equal to zero, a strategy outlined by [Yenier, 2017](#) to minimize M_L variations between different stations. In accordance with Equation (3.3), when we have a total of "p" earthquakes and "q" stations, the universal structure of the kernel matrix "G", along with the vectors "d" and "m".

4.4 Results and Discussion

Figure 4.2 shows the decrease in normalized Wood-Anderson (WA) amplitudes as a function of hypocentral distance within the three reference distance intervals. It is important to highlight that around 80 and 200 km, a noticeable change in amplitude decay is observed, which is believed to be due to the Moho-bounce effect. This phenomenon varies across regions and is influenced by multiple factors, including crust thickness, velocity structure, Moho boundary, dip angle, and regional anelastic attenuation, as discussed by ([Bakun and Joyner, 1984](#); [Kavoura et al., 2020](#)). It

occurs when reflected and refracted waves merge with direct waves, thereby altering the attenuation pattern at intermediate distances [Yenier, 2017](#). Similar findings have been documented for North America (± 70 and ± 150 km) ([Bakun and Joyner, 1984](#); [Burger et al., 1987](#); [Atkinson, 2012](#)), South Egypt (90-175 km) ([Abdullah et al., 2022](#)), and western Canada (100-220 km) ([Yenier, 2017](#)). To identify the transition distances, we conducted a grid search between 50 km and 150 km for R1 and between 100 km and 250 km for R2, with a 5 km increment. We also considered the possibility of a bilinear geometric spreading with $R1=R2$. Using Equation (3), we calculated the model coefficients ($n1$, $n2$, $n3$, b , S_j , and M_{Li}) and selected the best-fitting parameter set that resulted in the lowest mean absolute residuals of each R1-R2 combination. Using a trilinear function with the selected transition distances equal to $R1=90$ and $R2=190$ km, we calculate the correction distance term as follows:

$$-\log_{10}(A_0) = \begin{cases} 0.6747 * \log_{10}(R_{ij}) + 0.0002 * R_{ij} + 1.6306 & R_{ij} \leq 90 \\ 1.7736 * \log_{10}(R_{ij}) + 0.0002 * R_{ij} - 0.5169 & 90 < R_{ij} \leq 190 \\ 2.4580 * \log_{10}(R_{ij}) + 0.0002 * R_{ij} - 2.0765 & R_{ij} > 190 \end{cases} \quad (4.2)$$

Figure 4.3 shows a strong correlation between our $-\log_{10}(A_0)$ values and the dataset. Specifically, we note a greater attenuation in northern Algeria in comparison to other regions, as indicated by the higher $-\log_{10}(A_0)$ values observed for distances less than 30 km. We can compare these values with those obtained for the Ethiopian plateau ([Brazier et al., 2008](#)), especially in the hypocentral distance range of 30-90 km. The determined $-\log_{10}(A_0)$ value for the northern Algeria region in this study slightly deviates from values in Southern California. As a result, utilizing the relationship outlined in [Hutton and Boore, 1987](#) to calculate M_L in the region may lead to an underestimation of seismic magnitude, particularly for stations located at shorter distances. For distances between 90 and 600 km from the hypocenter, our attenuation curve approximates the characteristics found in southern California ([Hutton and Boore, 1987](#)) and Italy ([Di Bona, 2016](#)). In Figure 4.4, we present the residuals between event magnitudes (calculated as the average magnitude of all stations for M_{Li}) and station magnitudes (M_{Lij}), as a function of hypocentral distance. Unlike Figure 4.4b, Figure 4.4a incorporates the station correction in the magnitude calculation. Over most of the distance range, we observe nearly zero average residuals, which highlights the

efficacy of our computed distance correction in capturing the attenuation characteristics of northern Algeria events across distances that range from 5 to 600 km.

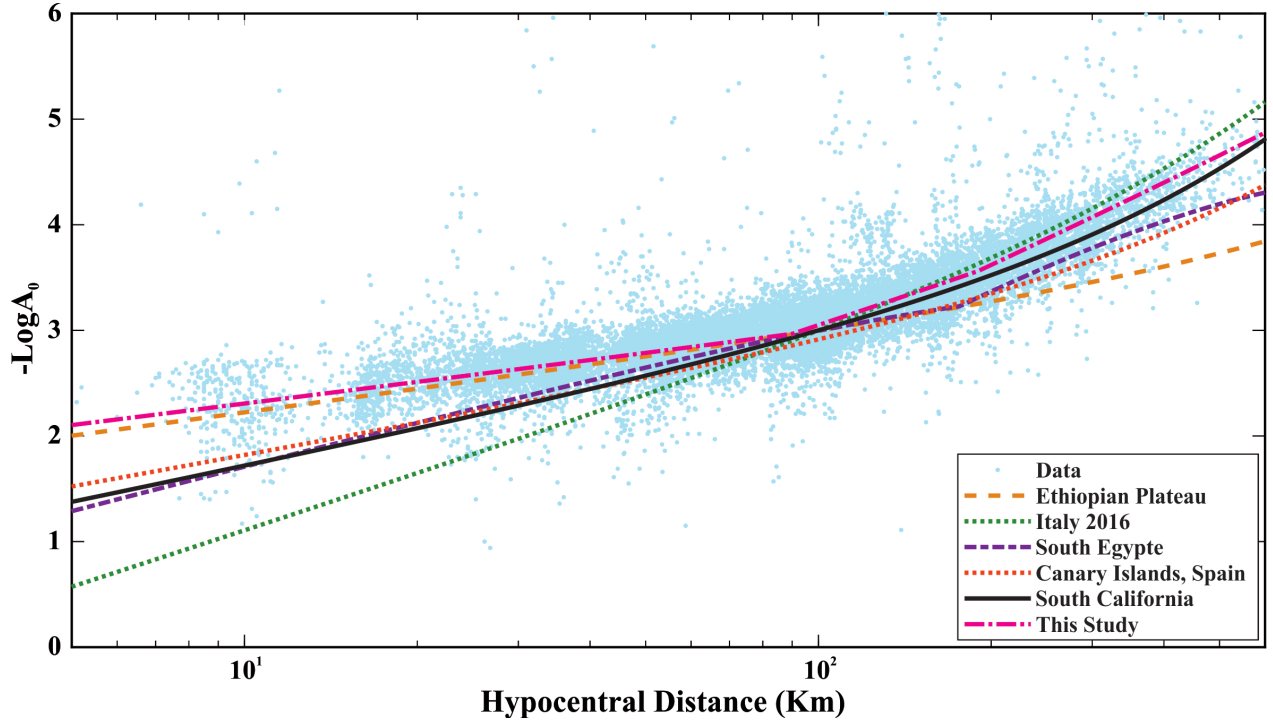


Figure 4.3: Comparison of the attenuation curve of the Northern Algeria region (Equation 6 of this study) with those of other regions, including Southern California (Hutton and Boore, 1987), the Ethiopian Plateau (Brazier et al., 2008), Italy (Di Bona, 2016), South Egypt (Abdullah et al., 2022), and the Canary Islands (Rueda and Mezcua, 2022).

Figure 4.5 displays the station correction terms 'S' for the 72 seismograph stations. These corrections are dependent on the ground conditions and instrument installation at each location, as established by Richter, 1958. It's crucial to note that a positive correction factor corresponds to a decrease in wave amplitude, while a negative value corresponds to an increase in wave amplitude. Our analysis indicates a range of correction factors for stations, ranging from -0.50 to 0.54, as illustrated in Figure 4.5. Among the 72 stations examined, approximately two-thirds have a low station correction factor ($-0.2 \leq S \leq 0.2$), while the remaining third have a high station correction factor ($|S| \leq 0.2$). These results imply that these stations may have poor coupling or are located in hilltops with complicated geological conditions (Saunders et al., 2013). Indeed, Layadi et al., 2023 have shown that 17 broadband ADSN stations installed in different geological and topographical contexts are not free from site-effects.

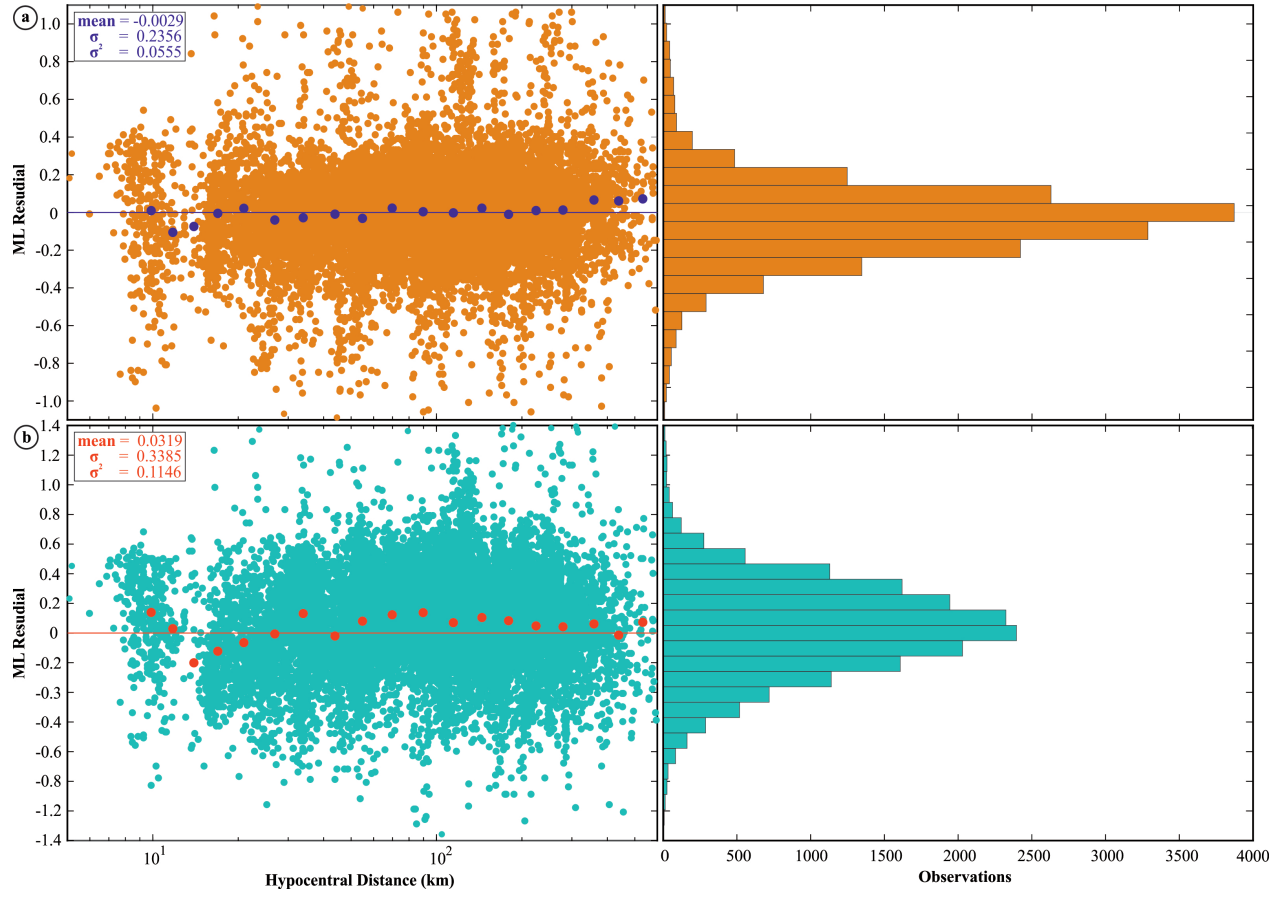


Figure 4.4: Local magnitude residuals (difference between the station and event local magnitudes), as a function of hypocentral distance, and histogram of the residuals. (a) The calculated magnitudes include the station correction term. (b) The calculated magnitudes do not include the station correction term. The red and the blue dots around the horizontal line represent the average residuals.

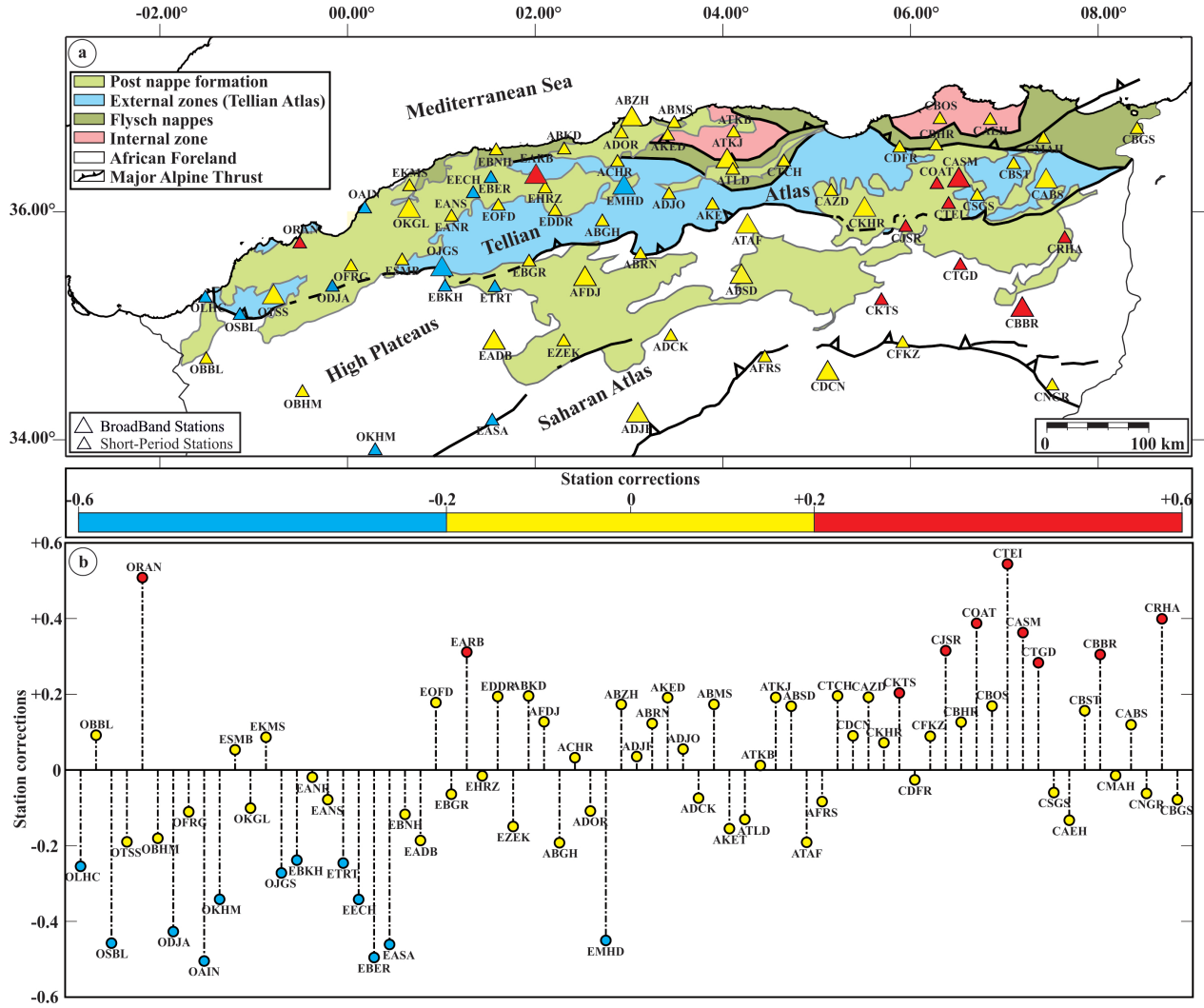


Figure 4.5: (a) Spatial distribution of station correction factors and their correlation with geological structures in the study area. (b) Colored circles indicate negative and positive correction factors for each station.

Figure 4.5a presents the correction factor values of the stations, displayed with a color scheme superimposed on a tectonic map of northern Algeria. The map gives an overview of the significant geological domains present in the area. The colors yellow, red, and blue represent low, high positive, and high negative correction factors, respectively. In Figure 4.5b, the exact values of these factors are provided for each station. Our analysis revealed that the large positive correction factors are concentrated in the eastern part of the region (domain 2), particularly in the Mila-Constantine basin (MCB, Figure 1.2) and the Aures Ranges. These areas are known for their heterogeneous soil composition and numerous fault segments with varying orientations, contributing to the attenuation

of seismic wave amplitudes at the stations installed there. However, stations in the far northeast exhibit low correction factors as they are located in the more homogeneous LKB region. It should be noted that the two stations EARB and ORAN, located in an area considered to be homogenous, have a high correction factor. This could be attributed to local site conditions or poor soil coupling. The central and western parts of the region (domain 1) are also relatively homogeneous, resulting in low correction factors for the majority of stations. However, our analysis also identified several stations with high negative correction factors located in three main regions: the Chellif basin, the Tellian Atlas Mountains, and the Saharan Atlas Range. These are likely due to the site effect condition, which could be caused by: (1) the sedimentary layers in the Cheliff and Oued Fodda basins as indicated by [Layadi et al., 2016](#) and [Issaadi et al., 2020](#) for the EECH and EBER stations respectively. (2) the topographic effect in the two other regions, including the Tellian Atlas Mountains like Dahra and Blida Atlas for the OAIN and EMHD stations respectively, and the along of Saharan Atlas for the EASA and OKMH stations. In general, there appears to be no significant correlation between station corrections and regional geology (Figure 4.5a). However, a minor correlation can be observed for the red stations, characterized by high positive corrections, predominantly situated on the eastern side. These stations are primarily located within the post-nappe formations of the Mila-Constantine basin. Unlike other intra-mountain basins within the Atlas Tellian, this particular basin is positioned at a higher elevation due to a subsequent compressive tectonic phase following its initial deposition ([Abacha, 2015](#) and related references). This geological context has led to the emergence of various mountain ranges (such as Chettaba, Fellten, Ouled Salem, etc.) and a network of active faults oriented in different directions (e.g., the North-Constantine Fault, Ain Smara Fault, or NCF, ASF as shown in Figure 1.2c). This geological heterogeneity could potentially influence the seismic attenuation within this basin. The positions of the OJGS, ETRT, EBKH, ODJA, OSBL, and OLHC stations show a clear alignment along the southern deformation front of Tell (Figure 1.2c). According to [Layadi et al., 2023](#), who only studied broadband ADSN stations, the OJGS station is affected by a topographic site effect. The same effect could also affect the other stations mentioned above, since they are installed on the same structure. At this stage, we do not have sufficient evidence to fully explain the alignment of these amplified stations along the southern deformation front of the Tellian Atlas. In particular, if we consider that other stations located on the same front did not require similar corrections.

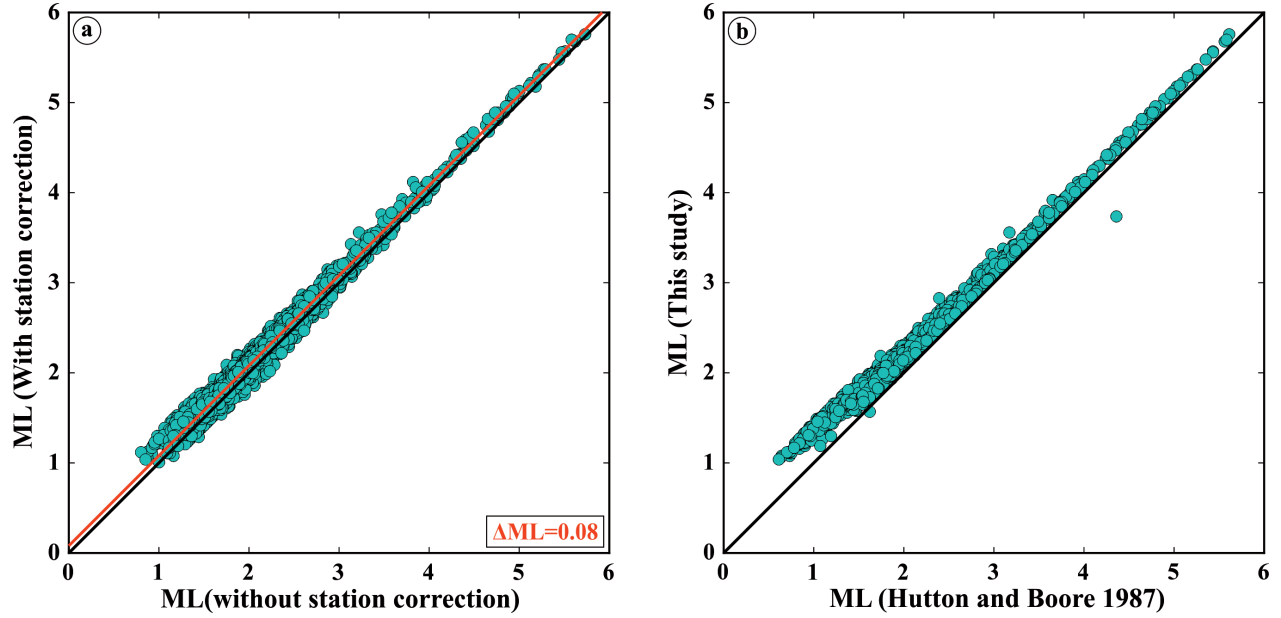


Figure 4.6: Comparison of average event magnitudes calculated for earthquakes in Northern Algeria. (a) Between the M_L scale derived in this study with and without station correction term. (b) Between the [Hutton and Boore, 1987](#) scale, and the M_L scale derived in this study. The red line represents the regression model, while the black line indicates the 1:1 relationship..

In Figure 4.6a, a comparison of estimated M_L magnitudes with and without the station correction factor is presented. On average, there is a difference of 0.1 units between the two sets of magnitudes across the entire dataset. Furthermore, Figures 4.4a and 4.4b illustrate station magnitude residuals with and without applied station correction. It is clear that incorporating the station correction reduces the average residuals in the dataset. After applying the station correction, the magnitude residuals display a standard deviation of 0.2356 and a variance of 0.0555. In contrast, without the station correction, the values increase to 0.3385 and 0.1146, respectively. The incorporation of the station correction yields a significant 30% decrease in the standard deviation, highlighting its importance in augmenting the precision of the newly derived scale and significantly reducing the variability of the residuals. A comparison was made between our calculated magnitudes and those obtained from the relationship of [Hutton and Boore, 1987](#), which was previously used to derive M_L in Algeria. Figure 4.6b shows the results. It is clear that the new M_L scale shows slightly larger values than those obtained with the Southern California magnitude scale. The average difference is about 0.2 units. The moment magnitude (M_w) is prioritized by scientists over local magnitude due to its greater accuracy in providing a more comprehensive measure of

earthquake size and energy release. Accurately estimating M_w for small-magnitude earthquakes is important, particularly in predicting ground motion at lower magnitudes. Many ground motion prediction equations rely on M_w as a key input parameter, making it common practice to establish empirical formulas that link M_w to locally used magnitude scales in seismology. In this context, we have derived an M_L - M_w conversion relationship based on a dataset comprising 55 events within the range of $2.8 \leq M_L \leq 5.8$ and $3.1 \leq M_w \leq 6$ which was derived from [Boulahia, 2022](#), using both moment tensor inversion and spectral analysis methods in the estimation of moment magnitude.

Utilizing linear orthogonal regression, we conducted a fitting analysis of the M_L - M_w scatter plot, yielding the equation $M_w = (0.8046 \pm 0.0271) * M_L + (1.0203 \pm 0.1155)$, characterized by a robust correlation coefficient of 0.93 and a standard deviation (σ) of 0.16. Similarly, [Bellalem et al., 2022](#) assembled a consistent M_w catalog for Northern Algeria and neighboring regions. Figure 4.7 visually presents the regression outcomes. Notably, [Bellalem et al., 2022](#) regression model aligns with our estimated 95% confidence interval. The disparity between their regression model and ours amounts to approximately 0.23 magnitude units. It is worth highlighting that [Bellalem et al., 2022](#) employed local magnitude values from the CRAAG catalog, which were calculated using the Southern Californian relationship, and the discrepancy between our magnitude values and the Southern Californian model is 0.2 units. Thus, the convergence of [Bellalem et al., 2022](#) regression model with our confidence interval is entirely expected and consistent.

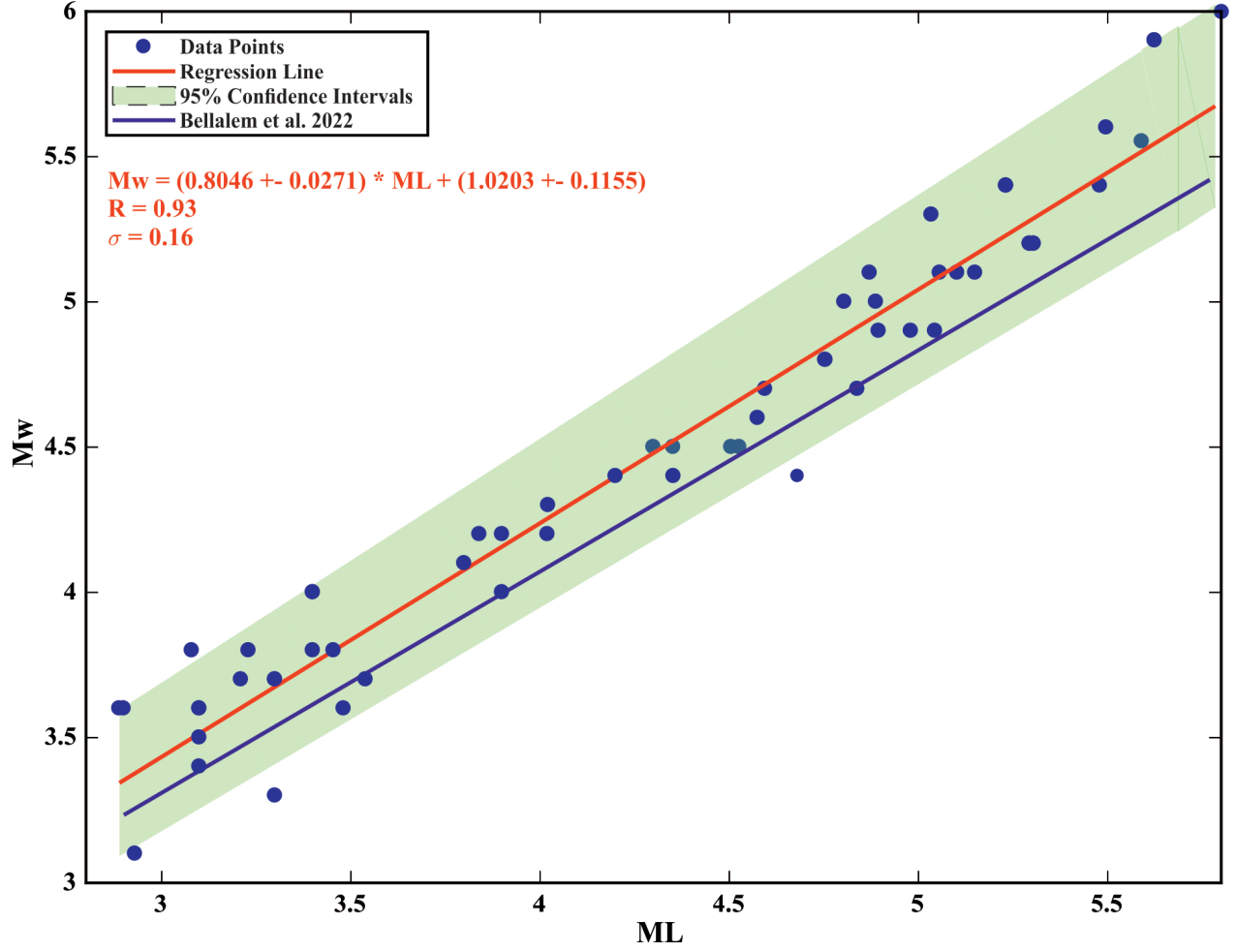


Figure 4.7: Relationship between M_w and M_L scales for Algerian earthquakes from [Roubeche et al., 2024](#). Blue circles represent the selected earthquakes, while the red and blue lines correspond to the regression models from this study and [Bellalem et al., 2022](#), respectively. The green shaded region indicates the 95% confidence interval.

4.5 Conclusion

This chapter presents a local magnitude (M_L) scale specific to northern Algeria. The M_L scale was derived from an in-depth analysis of seismic data comprising 1901 high-quality earthquake records. Analysis of amplitude decay over different distances in northern Algeria revealed the presence of two transition distances at 90 and 190 km, effectively dividing the region into three segments. To accommodate these variations, we determined the optimal distance correction term $-\log_{10}(A_0)$ that best fits our data. We found that the attenuation characteristics of seismic events in northern Algeria can be accurately described by a trilinear function, demonstrating its applicability

to the region. The distance correction formula we derived guarantees the accuracy of the M_L relationship over a distance range of 5 to 600 kilometers. A comparative analysis with other local magnitude relations, notably those for Southern California, has shown that our proposed M_L scale consistently produces slightly higher M_L values over all distances, with an average difference of 0.2 units. This underlines the unique seismic characteristics of northern Algeria and the need for a region-specific magnitude scale. In addition, by calculating the corrections for 72 stations, we effectively minimized the M_L residuals. These corrections highlight the influence of local site effects on the amplitude of the seismic signal. Finally, our study established an accurate, region-specific M_L relationship for northern Algeria, providing a valuable tool for seismic monitoring, hazard assessment, and earthquake research in the region. The knowledge gained from this work provides a better understanding of seismic activity in northern Algeria and improves the accuracy of magnitude estimates, contributing to the reduction of seismic risk and the protection of vulnerable communities

Chapter 5

A Duration Magnitude scale (M_d) for northern Algeria

5.1 Introduction

The study of earthquake hazards encompasses various strategies aimed at reducing the risks associated with seismic events. A key component of such studies is the understanding and quantifying the characteristics of earthquake sources. One of the most critical parameters in this context is the earthquake magnitude, a physical quantity that differentiates one seismic event from another. Magnitude is a fundamental measure of the size and energy released by an earthquake, making it indispensable in assessing the potential impact of seismic events and designing effective early warning system. By properly estimating the earthquake magnitudes, researchers improve the reliability of hazard assessment and enhance the earthquake mitigation. This highlights the importance of continuously refining methods for magnitude estimation and understanding its role within the broader framework of seismic hazard analysis.

The oldest earthquake magnitude scale, known as the local magnitude (M_L) scale, was introduced by [Richter, 1935](#). Over time, additional magnitude scales have been developed based on M_L to better characterize earthquakes. These include the surface wave magnitude (M_s), body wave magnitude (m_b), duration magnitude (M_d), and moment magnitude (M_w) scales, among others. Each scale considers different seismic parameters such as amplitude, duration, wave period, and wave type, providing complementary tools for magnitude estimation ([Havskov and Ottemoller,](#)

2010).

However, magnitude scales that rely on maximum amplitude, such as M_L , face limitations when measuring large-magnitude earthquakes or seismic events recorded at stations near the hypocenter. Additionally, accurate estimation of maximum amplitude is hindered when the instrumental response is unreliable or unavailable. In such cases, the amplitude of seismic waves often saturates or clips, leading to underestimation of the earthquake's true magnitude (Rodríguez-Pradilla and Eaton, 2019). To address this issue, the duration magnitude scale (M_d) is a good alternative. Unlike amplitude-based scales, M_d determines magnitude based on the seismic signal's duration or coda length, making it less prone to saturation and better suited for estimating magnitudes of large or nearby earthquakes.

Before calibrating the duration magnitude, it is essential to decide which magnitude scale to use and to verify that the data supports it. The current CRAAG catalog provides both M_L and M_w for moderate and large seismic events. Figure 4.7 illustrates the relationship between M_w and M_L for 58 earthquakes that will be used for calibration. Recognizing the inherent imprecision in obtaining M_w for small earthquakes, and considering M_L as a valid characterization of seismic events, we choose to use M_L in the calibration of the duration magnitude M_d .

The primary objective of this chapter is to estimate the duration magnitude scale (M_d) based on the local magnitude (M_L) of a series of seismo-tectonic events in northern Algeria. The study is based on a dataset compiled by Roubache et al., 2024, consisting of 1,901 earthquakes recorded at 72 stations in the Algeria Digital Seismic Network (ADSN) between January 2010 and June 2022. Initially, local magnitudes (M_L) were calculated from the dataset and subsequently analyzed to parameterize M_d values using the methodology described in Section 2.2.2. The chapter also incorporates an automated procedure for determining signal duration, offering a streamlined and efficient approach to M_d estimation using data from local seismic station networks around the epicenters.

This work lays the foundation for an upcoming publication. A key focus of the study is the development of a robust scaling relationship between M_d and M_L , advancing seismic magnitude estimation techniques and supporting applications such as earthquake early warning systems.

5.1.1 Data and Processing

In this part, we began by analyzing a dataset of 1901 earthquakes recorded at 72 stations of the ADSN network between January 2010 and June 2022. The events in this collection have magnitudes (M_L) ranging from 1 to 5.8. Initially, the signal-to-noise ratio (SNR) was calculated for each available seismogram to exclude records dominated by noise contamination from the analysis. The SNR was determined by comparing the pre-event noise amplitude to the maximum amplitude along the P-wave (Figure 5.1). We selected events recorded by at least three short-period stations with an SNR of 5 or higher and epicentral distances of less than 200 km. This criterion, determined through trial and error, effectively filtered out highly noisy data while retaining a substantial portion of the dataset. As a result, the dataset was refined to 11579 waveforms corresponding to 1872 events. The selected earthquakes have magnitudes ranging from 1.0 to 5.8, with epicentral distances spanning from 5 to 200 km. Figure 5.2 shows the epicenters of the selected earthquakes along with the stations used.

The duration (denoted as τ) is measured on the vertical component of ground velocity records. The method for calculating the duration of an earthquake event from seismic data is based on an approach developed by [Real and Teng, 1973](#), where the event duration is defined as the time interval between the arrival of the first P-wave (the fastest seismic wave) and the point where the signal coda (the tail end of the seismic signal) diminishes to the noise level, indicating that the seismic event has effectively ended. To automate the process of estimating this duration magnitude, we developed an algorithm. This algorithm measures the signal duration by analyzing the signal-to-noise ratio (SNR) within a moving time window as it progresses along the seismic record. This technique enables a fast and simple calculation of the event duration without manual intervention. Figure 5.3, provides a visual representation of the steps involved in this automated procedure. Initially, the data are preprocessed by removing the mean value and any linear trends to ensure that the signal represents the seismic activity accurately. An eight-pole, noncausal Butterworth band-pass filter with a frequency range of 1–20 Hz is then applied to the data to reduce contamination from high-frequency noise, as shown in Figure 5.3a. Next, the signal envelope is calculated by taking the square root of the sum of the squares of the original signal ($x(n)$) and its Hilbert transform ($y(n)$), depicted in Figure 5.3b. This envelope highlights the signal's amplitude variations. Finally, a moving average window is used to smooth the signal, making it clearer and more suit-

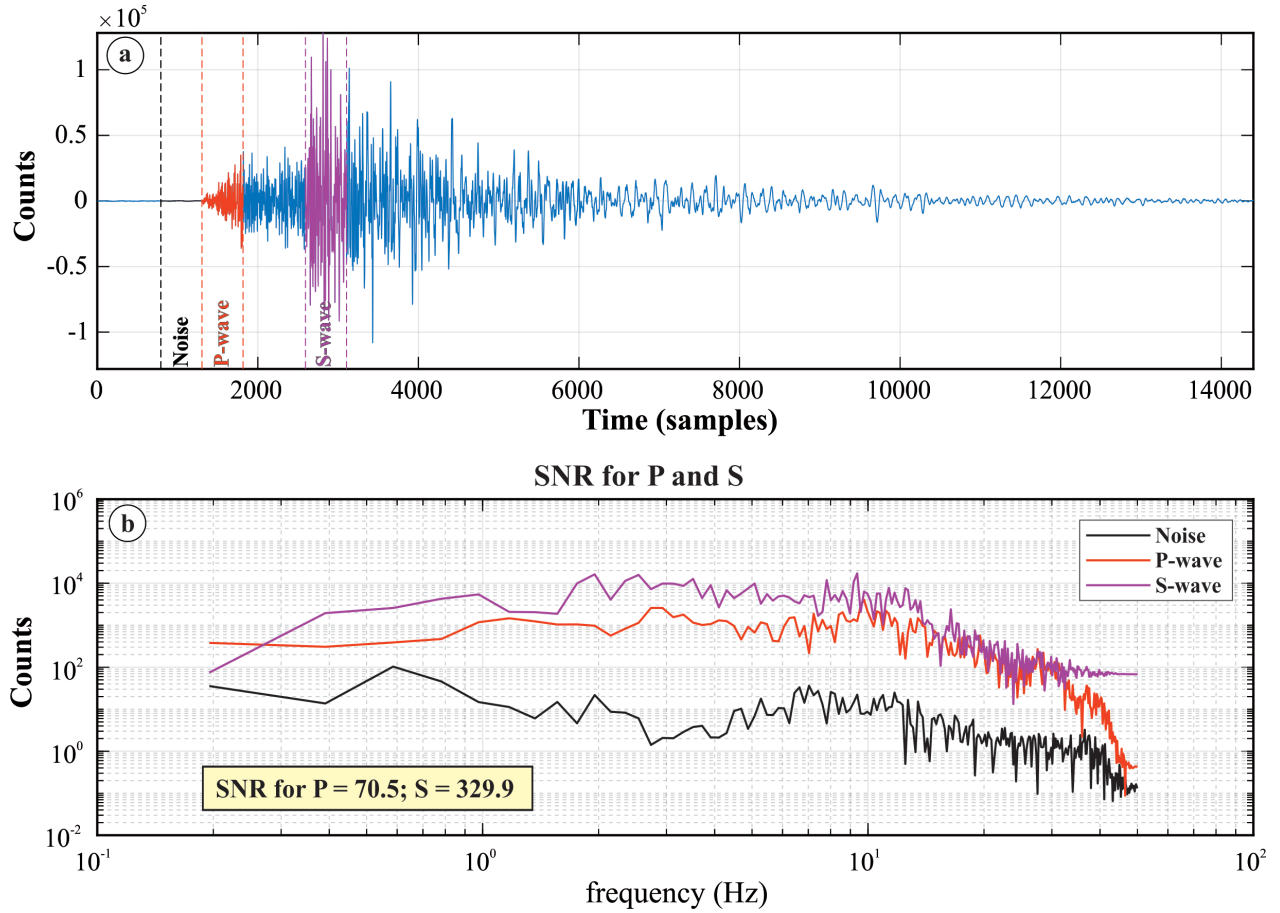


Figure 5.1: Illustrated here is an example showcasing the method for accounting for signal-to-noise ratio (SNR). (a) showcases selected windows depicting noise, P-wave, and S-wave. (b) presents the corresponding SNR results for both P- and S-waves.

able for subsequent processing, as illustrated in Figure 5.3c. The signal is considered to end when its amplitude becomes comparable to the pre-event noise level, which occurs when the following condition is met:

$$A_{Signal} \leq 1.05 * A_{Noise} \quad (5.1)$$

Here, A_{Signal} represents the average signal amplitude within a 0.5-second window, while A_{Noise} corresponds to the noise amplitude measured before the P-wave picking.

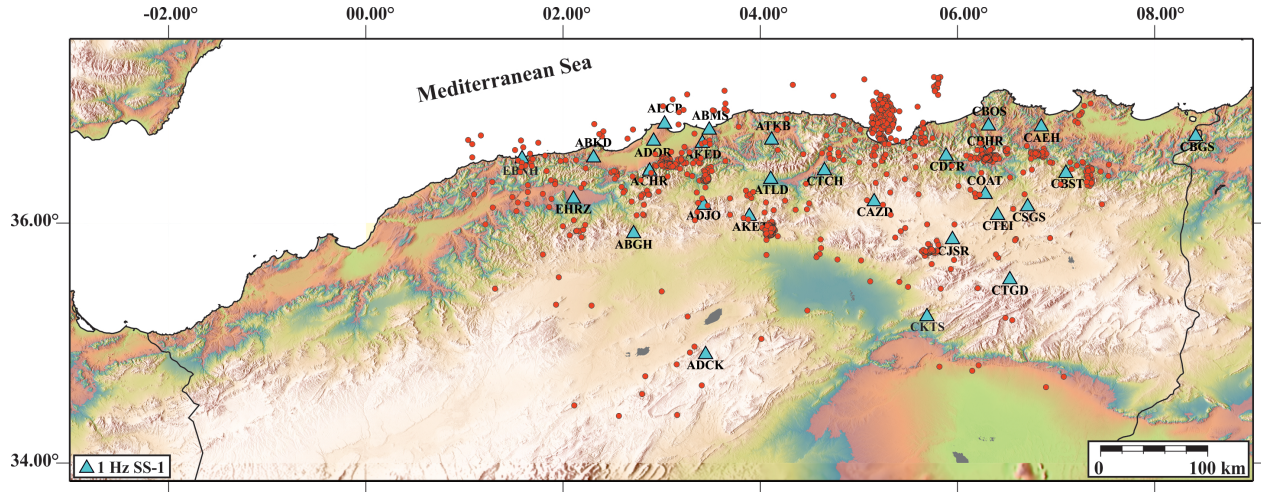


Figure 5.2: Illustrated here is an example showcasing the method for accounting for signal-to-noise ratio (SNR). (a) showcases selected windows depicting noise, P-wave, and S-wave. (b) presents the corresponding SNR results for both P- and S-waves.

5.2 Methods

5.2.1 Regression Analysis

The duration magnitude method (M_d) is one of the simplest techniques for estimating the magnitude of a local earthquake (Havskov and Ottemoller, 2010). As outlined by Herrmann, 1975, Real and Teng, 1973, this method involves defining a linear relationship between the signal duration and a known magnitude, expressed as follows:

$$M_d = a + b * \log(\tau) + c * R + S \quad (5.2)$$

Here, M_d represents the duration magnitude, τ is the signal duration in seconds, and R is the epicentral distance in kilometers. The constants a , b , and c are determined, while S is the station correction coefficient. Several studies (Del Pezzo et al., 2003; Castello et al., 2007; Bindi et al., 2005) have demonstrated that the distance-dependent term (c) contributes minimally to the duration magnitude for small hypocentral distances ($R < 150\text{km}$). A preliminary analysis of events located within and around the ADSN network, with a maximum epicentral distance of approximately 200 km, confirmed that the " c " coefficient in equation (5.2) is negligible. Consequently, we set $c = 0$ for the subsequent analysis. As mentioned in the previous section (section 5.1), the duration magnitude

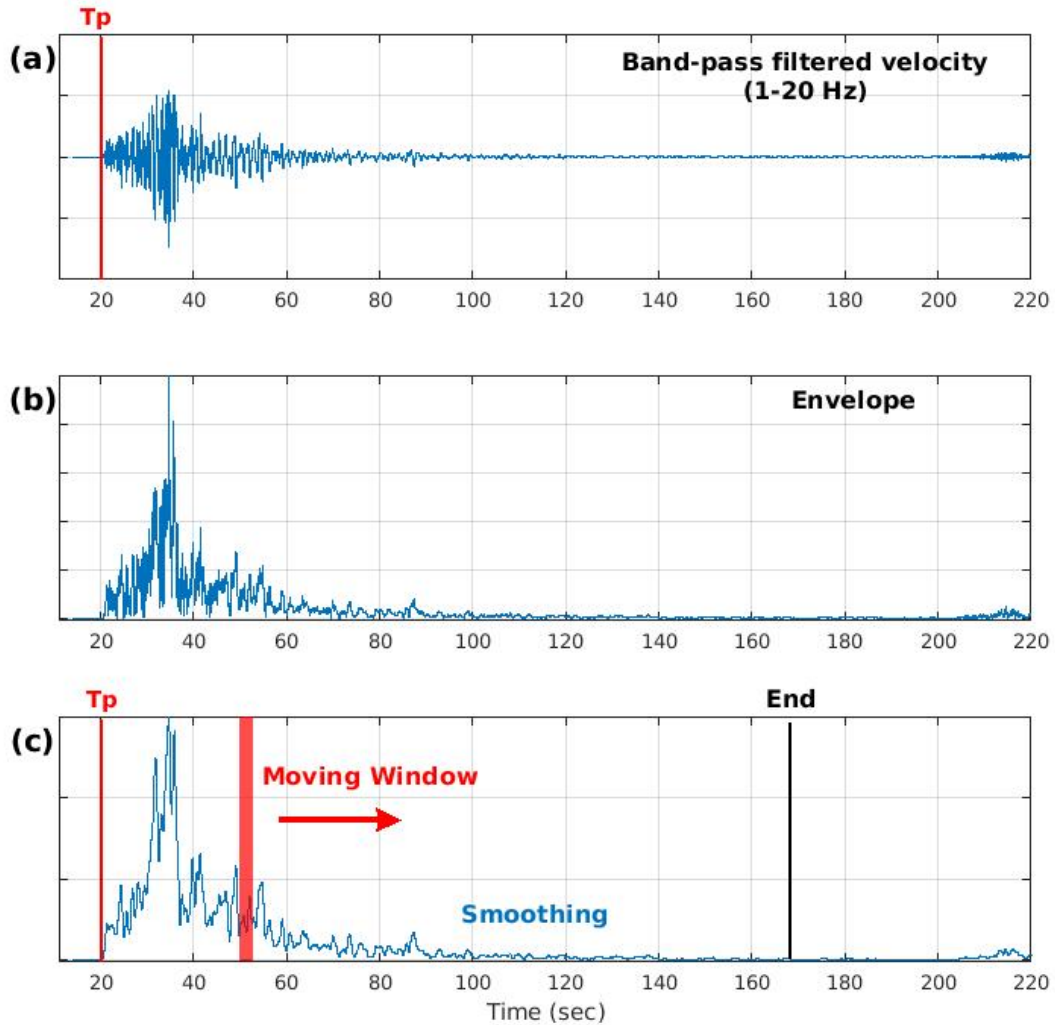


Figure 5.3: Sketch illustrating the procedure for the automatic measurement of signal duration. (a) The vertical component of a velocity record is processed, including mean removal, linear de-trending, and band-pass filtering using a noncausal eight-pole Butterworth filter in the 1–20 Hz frequency range. (b) The signal envelope is computed (details provided in the text). (c) The resulting signal in (b) is smoothed, and the begin and end markers are identified.

scale will be calibrated using the local magnitude (M_L) as a reference, assuming $M_L = M_d$ for each analyzed earthquake, the equation 5.2 became:

$$M_L = a + b * \log(\tau) + S \quad (5.3)$$

We measured the signal duration on each vertical velocity record using the automated procedure described above and analyzed its correlation with the local magnitude of the corresponding earthquake. The histogram in Figure (5.4) reveals that the data distribution across the magnitude range is uneven, with a notable dominance of low-magnitude data. This imbalance could bias the regression process, resulting in an underestimation of the slope parameter for the best-fit line and a low correlation coefficient. To mitigate this issue, we divided the dataset into magnitude classes with a width of 0.5, spanning the magnitude range from 1 to 5.8. One way to address Equation (5.3) is by expressing it as a system of linear equations in standard matrix form and solving the linear inversion problem $\mathbf{d} = \mathbf{G} \cdot \mathbf{m}$. Here, the vector $\mathbf{d}$ represents the known terms, specifically M_d , while the kernel matrix $\mathbf{G}$ defines the relationship between $\mathbf{d}$ and the model parameters $\mathbf{m}$, which include the unknowns such as a and b .

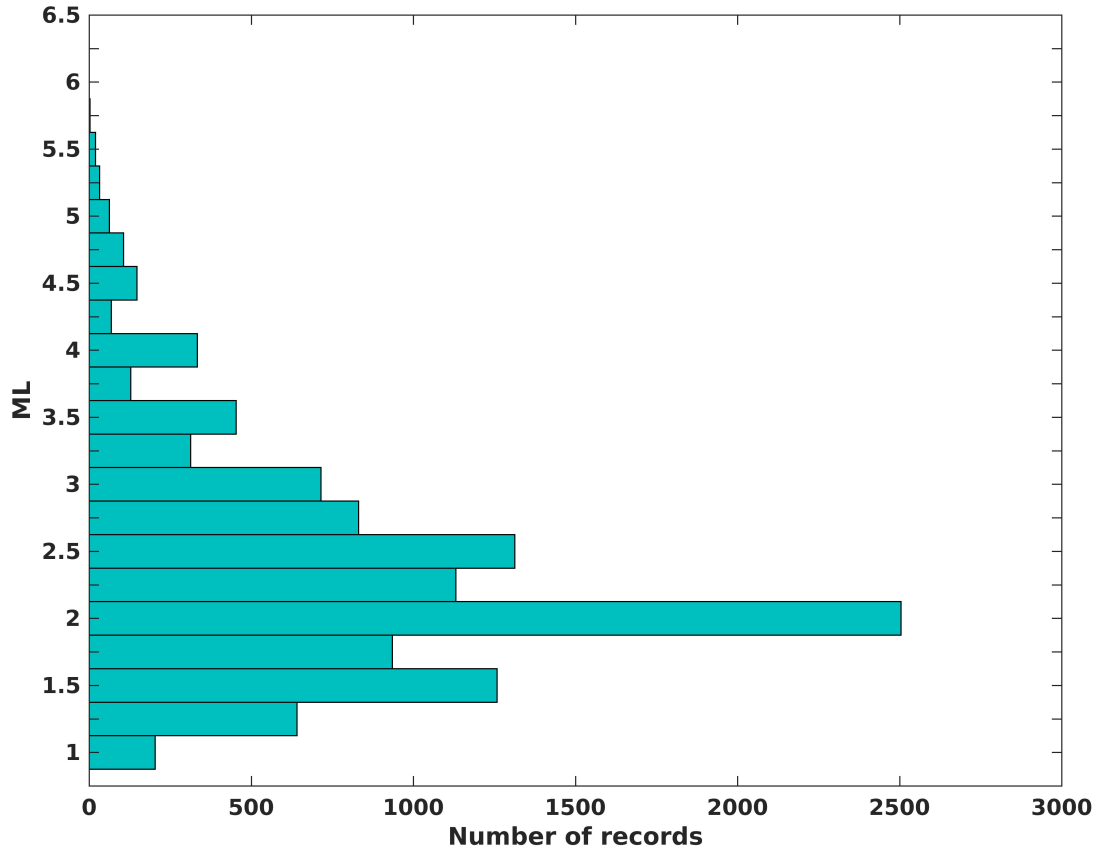


Figure 5.4: The distribution of records used in this study as a function of local magnitude shows that most of the data fall within the range $1.25 < M_L < 3$.

5.2.2 Station Corrections

The geological conditions of the propagation medium near recording station sites significantly affect the characteristics of the recorded seismic signal, including the length of its coda (τ). These site-specific conditions introduce systematic biases in the measurement of signal duration, leading to errors in the estimation of duration magnitude (M_d). To improve the accuracy of magnitude estimates, station correction coefficients (S_C) are determined for each recording station, as suggested by [Colombelli et al., 2014](#). These coefficients account for local site conditions, such as geological and soil properties, topography, ambient noise sources, and the coupling of the station to the ground, which can vary significantly between locations [Mushtaq et al., 2019](#). Station correction coefficients are derived by comparing observed and predicted magnitude values. For the i -th

seismic event and the j -th recording station, the station correction coefficient (S_j) is expressed as:

$$S_j = \frac{1}{N_j} \sum_{i=1}^{N_j} (M_{Li} - Md_{ij}) \quad (5.4)$$

Here, N_j represents the number of seismic events recorded at the j -th station, M_{Li} is the local magnitude for the i -th earthquake as provided by [Roubeche et al., 2024](#), and Md_{ij} is the duration magnitude estimated for the i -th event at the j -th station. This correction value is incorporated into the magnitude equation to minimize the overestimation or underestimation of the magnitude at a given station.

A positive or negative station correction coefficient primarily reflects the influence of site conditions, topography, and lateral velocity variations associated with heterogeneous near-surface structures ([Colombelli et al., 2014](#); [Joshi et al., 2020](#)). Seismic stations located on hard or old compact rocks, characterized by high acoustic impedance (i.e., high density and velocity) and low amplification, typically yield negative or near-zero correction coefficients. These conditions tend to underestimate the earthquake duration magnitude (M_d) due to the efficient transmission of seismic waves with minimal dissipation, resulting in shorter signal durations [[Mushtaq et al., 2021](#)].

Conversely, stations installed on loose materials, weathered rocks, young rocks, or soil characterized by low acoustic impedance (i.e., low density and velocity) generate positive correction coefficients, leading to overestimation of M_d . In such media, seismic signals undergo significant reverberation, amplification, and attenuation, with strong energy dissipation, producing longer signal durations and higher frequency content. These effects have been extensively documented ([Michaelson, 1990](#); [Colombelli et al., 2014](#); [Joshi et al., 2020](#); [Rueda and Mezcua, 2022](#)).

5.3 Results and Discussions

The duration τ , starts at the onset of the P-wave arrival and ends when the coda disappears into the background noise. This interval was measured automatically for seismic waveform data obtained from 28 short-period seismic stations of the ADSN, covering the period from June 2010 to June 2022. The duration measurements were performed on filtered records. The local magnitude (M_L) for each selected earthquake was used as a reference to calibrate the duration magnitude scale. The

results for the entire dataset are displayed as blue dots in Figure 5.5. Although a clear correlation between signal duration and earthquake magnitude is evident, the data exhibit substantial scattering, which is primarily due to the natural variability of duration when measured at different stations for the same earthquake. Additionally, the data are not uniformly distributed across the considered magnitude range, as indicated by the histogram in Figure 5.4. As previously noted, this uneven distribution suggests that the regression procedure applied to the entire dataset could be skewed by low-magnitude data, potentially leading to an underestimate of the slope parameter of the best-fit line and a poor correlation coefficient. To address this, we divided the dataset into magnitude classes with a width of 0.5, spanning the 1–6 magnitude range. For each magnitude class, we calculated the average duration, with the results shown as black circles in Figure 5.5. A similar approach has been employed by Colombelli et al. [2014], for example. The following relations were obtained by performing a linear regression of local magnitude (M_L) versus coda duration (τ):

$$M_d = (4.3621 \pm 0.1445) * \log(\tau) + (-4.9395 \pm 0.2925) + S_j \quad (5.5)$$

The station correction coefficients are obtained by comparing predicted and observed magnitude values. For the i -th event and the j -th recording site, the correction coefficients are calculated following the equation 5.4.

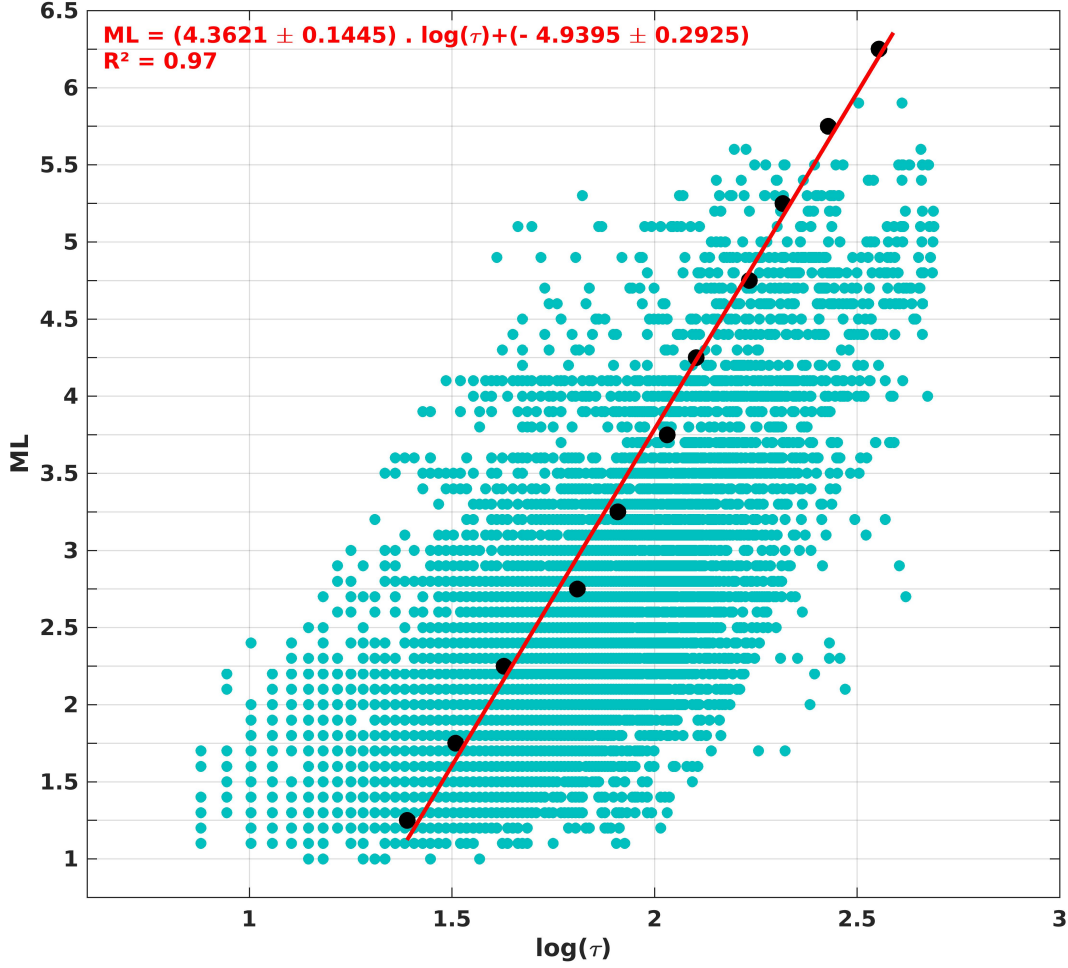


Figure 5.5: Local magnitude (M_L) as a function of the logarithm of duration (τ) for all selected events is shown by blue circles. Black points represent the results of the binning procedure, where each point corresponds to the average duration for different magnitude classes with a 0.5 magnitude unit step. The red line represents the best-fit curve, and the equation of the best fit, along with the correlation coefficient and standard error values, are reported in the panel.

To evaluate the statistical significance of the station correction coefficients, a z-test was conducted to assess whether each station correction significantly deviates from the null hypothesis of a Gaussian distribution with zero mean and a standard deviation of 0.15, which corresponds to the standard error found in the duration magnitude estimates. For each j -th station, the test statistic z_j is calculated using the following formula:

$$z_j = \frac{S_j}{\sqrt{(\sigma_i^2 - \sigma_T^2)/N_j}}$$

Where:

- S_j is the station correction coefficient for the j -th station,
- σ_i^2 is the variance of the observed magnitudes,
- $\sigma_T = 0.15$ is the standard deviation assumed for the null hypothesis (corresponding to the standard error in the duration magnitude estimates),
- N_j is the number of available data for the j -th station.

Assuming a significance level of 5%, the null hypothesis is rejected for stations where $|z_j| > 1.96$, while it cannot be rejected for the remaining stations. This means that stations whose z_j values exceed 1.96 have statistically significant correction coefficients and are included in the final analysis. Conversely, stations with $|z_j| \leq 1.96$ are excluded, as their correction coefficients do not significantly impact magnitude estimation.

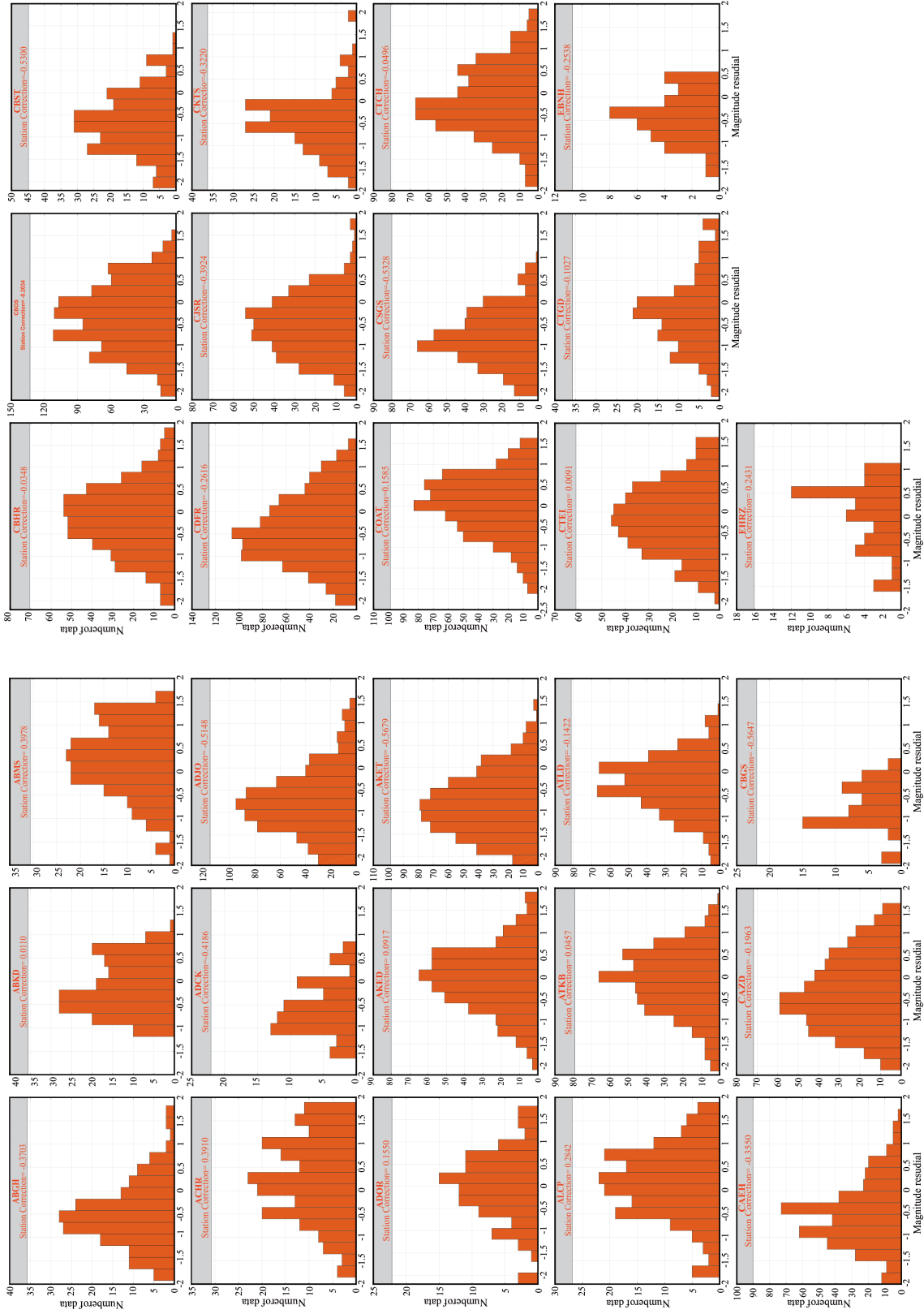


Figure 5.6: Distribution of magnitude residuals for each station.

Station coefficients, ranging from -0.57 to 0.39 (Figure 5.6), quantify the impact of local site conditions on recorded seismic signals. Negative coefficients are linked to areas with compact, old rocks that have high acoustic impedance. In these regions, seismic waves propagate efficiently with minimal energy loss, resulting in shorter signal durations and less scattering, which require negative corrections. On the other hand, positive coefficients are associated with young, unconsolidated sediments, often fractured and characterized by low acoustic impedance. These sediments amplify seismic waves and extend their duration due to reverberation and scattering, necessitating positive corrections. The farther a coefficient deviates from zero (+ direction), the greater the influence of local site properties on the seismic signal. This distribution of station coefficients offers important insights into the geological and geophysical characteristics of the areas surrounding seismic stations.

The analysis of residuals between M_d (duration magnitude) and M_L (local magnitude) values for all events reveals that the mean residual is approximately 0.2 (Figure 5.7). This result underscores the reliability of the M_d scale developed in this study for accurately estimating the size of seismic events recorded by the ADSN stations. The observed discrepancies between duration magnitude formulas across various regions, networks, and stations globally can be attributed to differences in attenuation, scattering, and station-specific conditions such as ambient noise levels (Havskov and Ottemoller, 2010). These variations highlight the importance of regionally tailored calibrations for magnitude scales.

The comparison between M_d and M_L demonstrates that event duration provides an independent and reliable estimate of earthquake magnitude. Given that the duration magnitude scale was calibrated under the assumption $M_d \approx M_L$ for each event, this outcome is not entirely surprising. A more robust evaluation involves comparing the duration magnitude (M_d) with the moment magnitude (M_w), as shown in Figure 5.8. An orthogonal regression analysis of M_d against M_w yielded the following relationship:

$$M_d = -0.87 \pm 0.09 + (1.29 \pm 0.05) \cdot M_w$$

The plot reveals a clear linear correlation between M_d and M_w . However, the best-fit regression line deviates significantly from the expected quadrant bisector, indicating that the relationship between these two scales is influenced by additional factors. This deviation may arise from regional

geological variations, source properties, or differences in how the two magnitude scales capture the characteristics of seismic events.

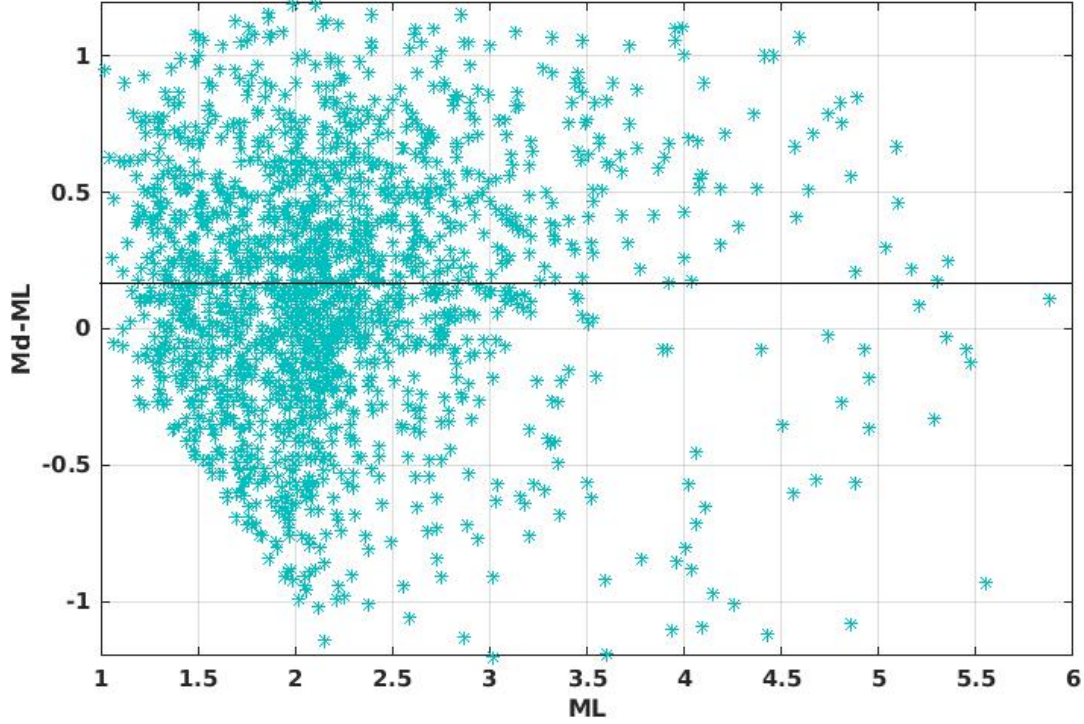


Figure 5.7: Duration magnitude residuals (difference between the duration and local magnitude), as a function of hypocentral distance.

The obtained results demonstrate a strong and statistically significant correlation ($R^2 = 0.66$) between M_d and M_L for magnitudes in the range of 1 to 6 and hypocentral distances of less than 200 km. This robust scaling relationship indicates that both magnitude scales provide consistent and homogeneous estimates for earthquakes recorded by the ADSN network. Furthermore, the relationship confirms that earthquake coda duration is a reliable metric for unbiased and accurate magnitude determination. This finding is consistent with the calibration process, where the M_d scale was derived under the assumption that $M_d = M_L$ for each earthquake, ensuring consistency between the two scales.

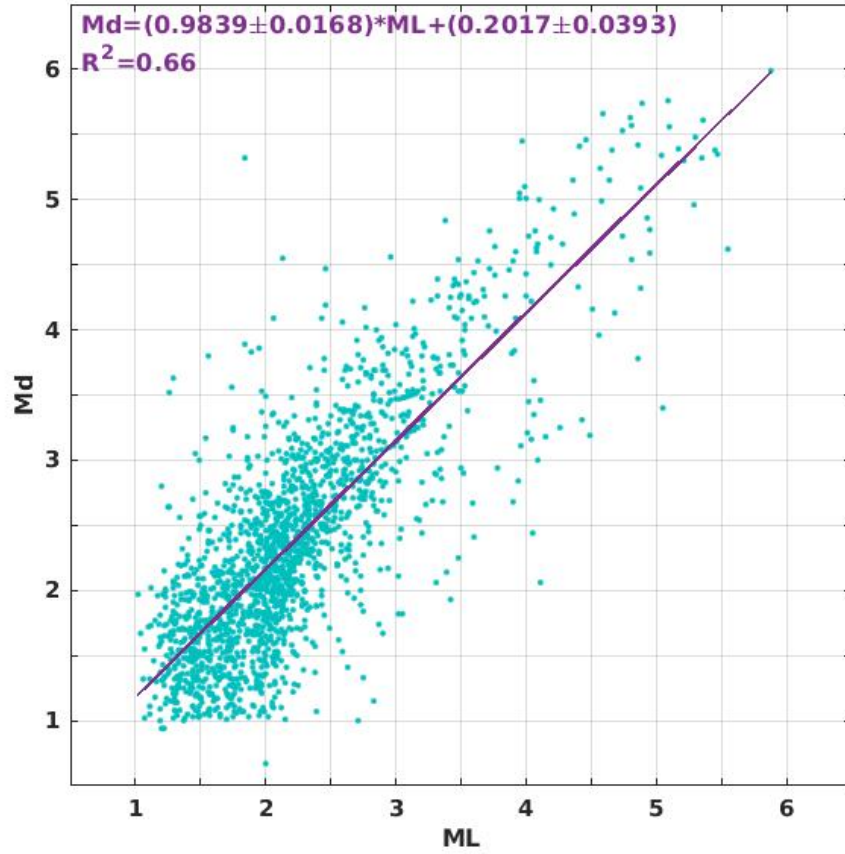


Figure 5.8: Relationship between M_d and M_L scales for Algerian earthquakes from [Roubeche et al., 2024](#). Blue circles represent the selected earthquakes, while the line corresponds to the regression models from this study.

Figure 5.9 presents a comparison between the duration station corrections obtained in this study and the amplitude station corrections previously reported by [Roubeche et al., 2024](#). This comparison aims to investigate potential correlations or patterns between the two types of station corrections. Identifying such relationships is crucial, as it may provide insights into the influence of local site effects, station-specific noise characteristics, or other geophysical factors on the measurement and calibration of seismic parameters. By examining these corrections, we can better understand how station-dependent adjustments affect the accuracy and consistency of magnitude estimations across the network.

Figure 5.9 demonstrates that the data points from ADSN stations generally deviate from the line representing equality between amplitude and duration corrections (the correction more than

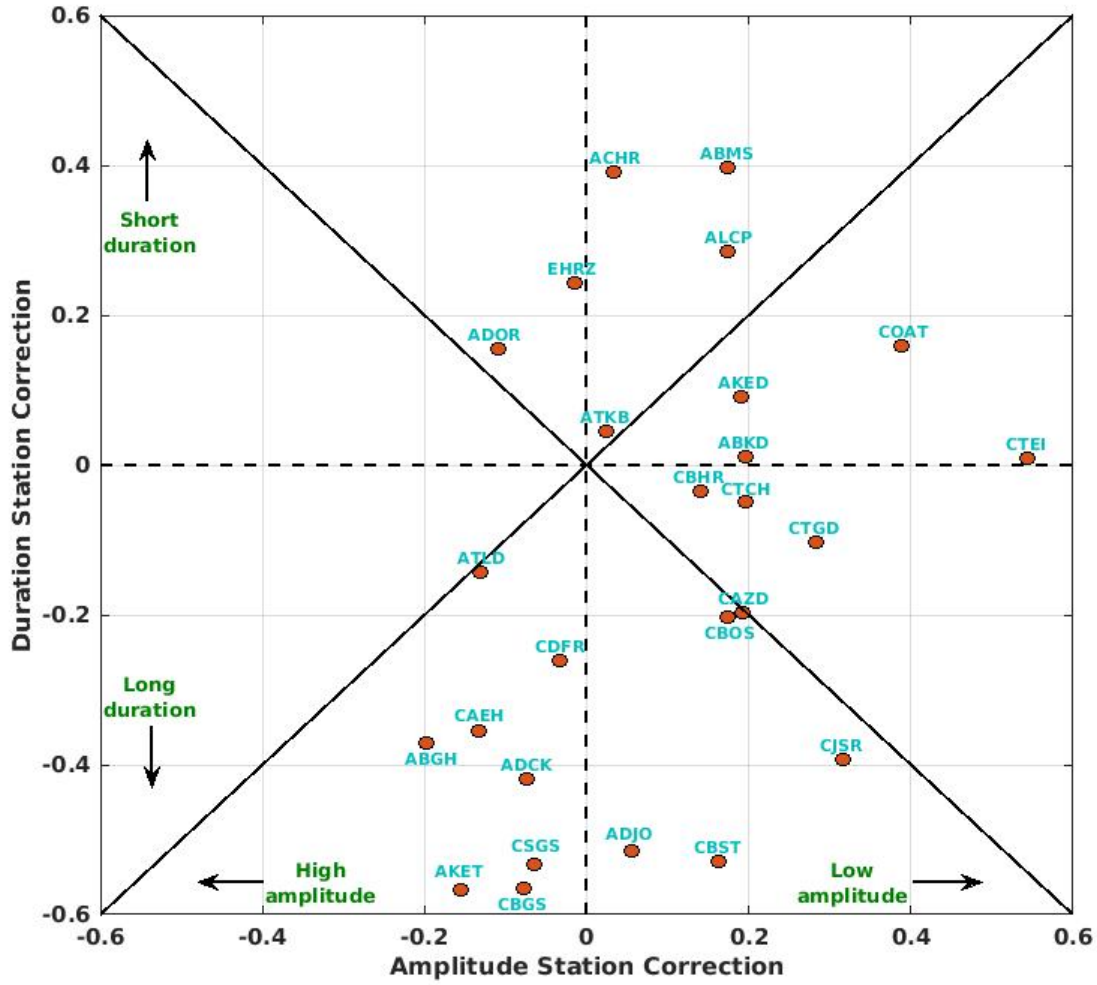


Figure 5.9: Plot of amplitude and duration corrections for the ADSN stations.

$|S|=2$), with a few exceptions. Negative corrections are observed for stations ABGH, AKET, CSGS, CBGS, and ADCK, located at elevations of 999 m, 1240 m, 967 m, 1219 m, and 1200 m, respectively. These corrections are attributed to the relatively high amplitudes and extended durations characteristic of the seismograms recorded at these stations. For this scenario, it is challenging to establish a consistent correspondence between the amplitude and duration corrections, as all stations exhibit negative duration corrections. Notably, for stations CJSR, CBST, and ADJO, the amplitude corrections are positive (indicating lower amplitudes), whereas the duration corrections are negative (indicating longer durations). This observation suggests that large amplitudes on seismograms do not necessarily correspond to prolonged coda durations at these stations. The dis-

crepancies in the responses of these sites underscore the significant influence of topographic and geological characteristics on seismic signals, which likely vary across the network. These factors introduce variability in how seismic waves are recorded, emphasizing the need for site-specific considerations in seismic analyses.

Application of Duration Magnitude for Earthquake Analysis

In this part, we present a straightforward and intuitive application of the previously derived duration magnitude (M_d), which relies on the simultaneous measurement of amplitude and signal duration. The example provided serves as a case study to explore the limitations and applicability of the adopted (M_L) and proposed (M_d) magnitude scales. It is not intended as a generalized or rigorous approach, as further testing across a broader range of earthquake magnitudes and distances would be necessary to establish a reliable methodology.

The core idea is derived from the well-established dependence of amplitude and signal duration on the source-to-receiver distance. Typically, for an earthquake of a given magnitude, the closer it is to a recording station, the smaller the P-S travel-time difference and, consequently, the shorter the signal duration at the station. Conversely, for distant sources, the increased S-P delay time and the arrival of later, indirect phases lead to a longer signal duration.

As an example, consider two earthquakes:

- **Tamanrasset Earthquake (19/10/2024):** This event was recorded only by the TAM station of the GEOSCOPE network due to the TTAM station which was not operational at the time of the event, and the nearest ADSN stations are more than 1300 km from the epicenter. Furthermore, this event hasn't been repertoried by any of the reputed seismological centers (e.g., NEIC, IGN, INGV...). Consequently, and knowing that locating earthquakes using a single station inherently carries high uncertainty, but at least two parameters of interest can be deduced using records from a single-station: event azimuth and epicentral distance.
- **Beni Ilmane Earthquake ($M_L = 3.9$, [Roubeche et al., 2024](#)):** This event occurred in the Beni Ilmane region and was recorded by the ADSN station, approximately 235 km from the epicenter.

Figure 5.10 illustrates the vertical component waveforms recorded by TAM and CSGS, respectively. Using the automatic procedure described earlier, the signal envelope for both records

is shown in Figure 5.10, where markers T_1 and T_2 indicate the beginning and end of the signal. Interestingly, despite the differing source-to-receiver distances, the distant and local earthquakes exhibit similar durations (~ 145 s and ~ 134 s, respectively).

This observation suggests that signal duration alone may not reliably distinguish between near and far events, as large durations can also result from nearby, larger earthquakes where ground motion amplitude takes longer to attenuate. However, combining duration with amplitude offers a solution:

- **Large duration + large amplitude:** Likely indicates a nearby earthquake.
- **Large duration + small amplitude:** Suggests a distant event.

Using Equation 5.5, we estimate the duration magnitude for these events:

$$M_d = 4.35 \quad (\text{Tamanrasset})$$

$$M_d = 3.9 \quad (\text{Beni Ilmane})$$

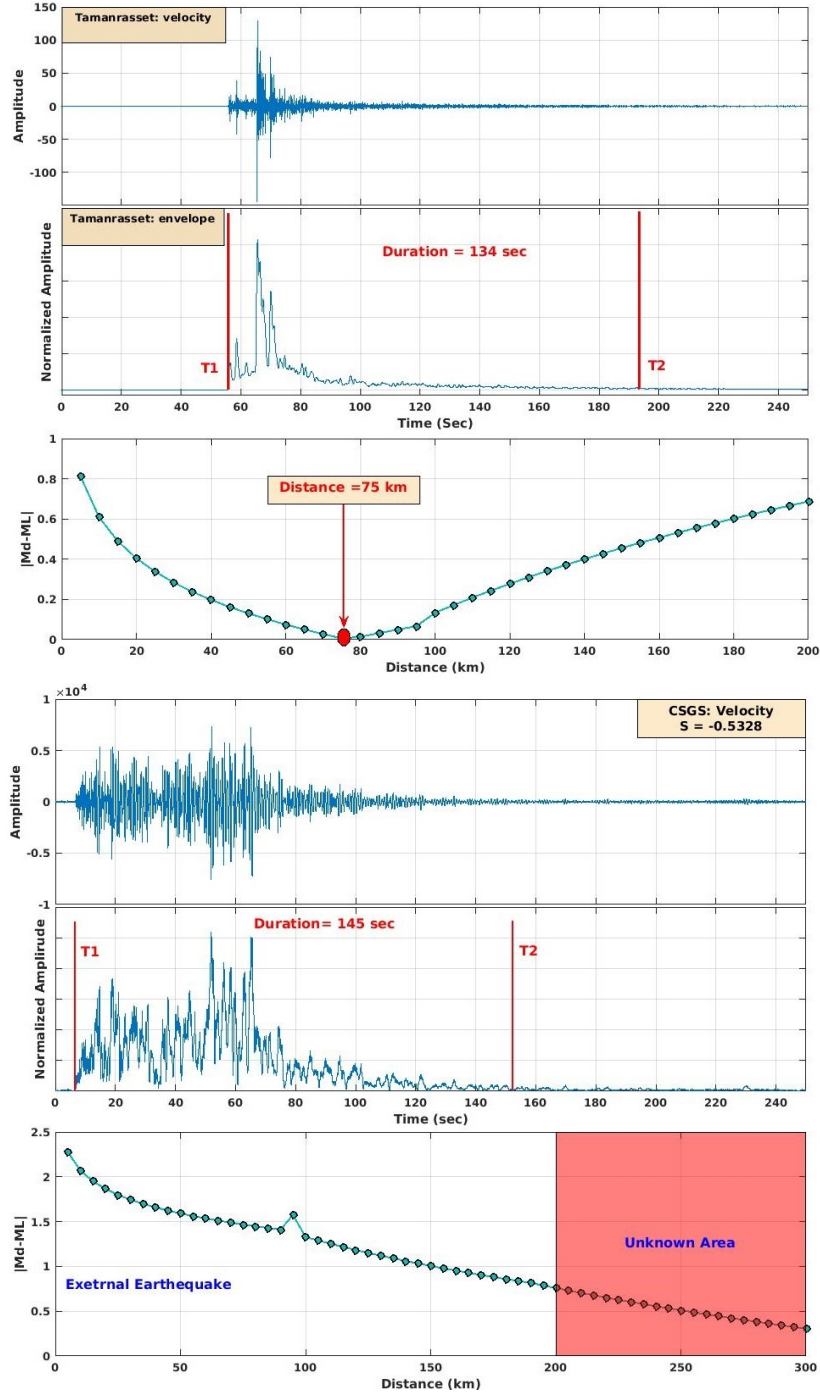


Figure 5.10: Analysis of magnitude residuals for each station. The first three panels display the raw signal, the duration estimated from the envelope, and the magnitude residuals, respectively, for the Tamanrasset event. The last three panels present the same analysis for the Beni Ilmane event.

For the Beni Ilmane event, the calculated M_d aligns closely with M_L ($M_d = 3.98$ calculated using all available stations), reflecting the calibration of the M_d relationship for events within the

northern Algeria region. For Tamanrasset, the underestimated M_d was expected, as Equation 5.5 does not explicitly account for distance dependency.

To estimate local magnitude (M_L) for Tamanrasset, we test distances ranging from 5 km to 200 km and compute the corresponding M_L values using the relationship in Roubeche et al., 2024. The comparison between M_d and M_L (Figure 5.10) reveals that discrepancies are significant when inaccurate distances are assumed. As testing distances approach the true value, residuals ($|M_D - M_L|$) decrease.

For Tamanrasset, the residual curve reaches a minimum around 75 km, suggesting this distance best explains the observed duration and amplitude. For Beni Ilmane, residuals decrease with distance but do not reach a minimum within 200 km, implying a more distant source.

The M_d scale was calibrated for earthquakes within Northern Algeria at distances up to ~ 200 km from ADSN stations. Beyond this range, M_d estimates are less reliable. Nevertheless, the trend of magnitude residuals can provide qualitative insights into source distance:

- **Increasing residuals ($M_d - M_L$) with distance:** The event is likely close to the station, and the minimum residual point provides an approximate distance.
- **Decreasing residuals ($M_d - M_L$) with distance:** The event is likely far, but the exact distance cannot be reliably estimated.

This qualitative approach offers a rapid, rough estimate of relative source-to-station distance and can help distinguish between events inside or outside the network (distance >200 km). With multiple stations, an automated decision scheme could be implemented, enabling rapid discrimination of internal versus external events and valuable information for automatic earthquake bulletins.

5.4 Conclusion

In this chapter, we developed a tool for the automatic measurement of signal duration from seismic records and calibrated a duration magnitude (M_d) scale for the Algerian Seismic Network in northern Algeria. The algorithm for estimating signal duration relies on the evaluation of the signal-to-noise ratio (SNR) within a moving time window, combined with a comparison between

the event's amplitude and the pre-event noise level. The M_d relationship was derived through linear regression analysis, with the data grouped into distinct magnitude classes. For each station in our database, station correction coefficients were computed, and their statistical significance was assessed using a z-test. After calibrating the M_d scale, we compared M_d estimates to both the local magnitude (M_L) and moment magnitude (M_w) values, as reported by [Roubeche et al., 2024](#). The comparison revealed excellent agreement between M_d and M_L values and a clear linear correlation between M_d and M_W , confirming that event duration provides a robust, independent estimate of earthquake size. The M_d relationships established in this study allow for the computation of duration magnitudes within the range of 1 to 6 for earthquakes in northern Algeria. These relationships account for regional geological variations, source-to-station distances, and site-specific geological characteristics at the recording stations.

The duration magnitude scale developed here complements the local magnitude scale previously proposed for the region, offering an additional reliable metric for earthquake size estimation. Both scales yield consistent and homogeneous magnitude values for seismic events in northern Algeria, thereby enhancing the capability of the Algerian Seismic Network to characterize seismicity in the region.

General Conclusion

The initial objectives of this thesis focused on the estimation of local and duration magnitudes for northern Algeria. This region offers a good study area due to the existence of high-quality earthquake data over an acceptable period (~ 20 years) obtained by the ADSN. The Algerian catalogs already contain moment magnitude (M_w) for some events, and duration magnitude (M_d) values, providing a solid foundation for further seismic research. CRAAG has also worked on creating a uniform duration magnitude catalog to facilitate the estimation of unbiased earthquake recurrence rates in northern Algeria, which is crucial for seismic hazard assessments.

However, despite these advancements, the magnitude scale relationships used in creating the catalog, especially for local and duration magnitudes, are not entirely suitable for northern Algeria. This is due to the region's unique seismic and geological characteristics, which influence attenuation properties. Additionally, the lack of sufficient data (until now) has also contributed to this limitation. The development of reliable empirical relationships for magnitude estimation is currently hindered by significant challenges, including data heterogeneity, biases, and systematic errors. Magnitude relationship approaches, and geological conditions further complicate the process of establishing consistent and accurate empirical relationships.

A key issue arises from biases and systematic errors in the magnitude values calculated by local networks. If local magnitude values are affected by uncorrected biases or errors, integrating them into empirical relationships can propagate inaccuracies, ultimately leading to unreliable outcomes. This underscores the importance of addressing these issues in the development of a more constrained local magnitude scale.

In this thesis, we estimated the local magnitude (M_L) scale for northern Algeria, motivated by the need for a scale specifically adapted to the region's seismic characteristics. This marks the first M_L scale developed for northern Algeria, offering more accurate magnitude estimates than

those produced by previous scales (unpublished scales). Indeed, previously the "service technique" estimates M_d using an unpublished scale. ADSN used (still using) a modified (Hutton and Boore, 1987) relationship for local magnitude estimation, which relied on a distance correction model that may not be suitable for the seismic conditions of northern Algeria. Our analysis revealed a difference of 0.2 units in the magnitude estimation when using this model. To address this discrepancy, we performed a constrained parametric inversion and identified a more complex distance correction specific to the region.

Further validation was conducted by comparing our newly developed M_L scale to a reliable MW solution. Our M_L values showed better agreement with M_w than the previous M_L values, confirming that the Southern California-based distance correction was not adequate for the attenuation properties of northern Algeria. Additionally, we derived station corrections and recalculated M_L for 1,901 earthquakes, improving the accuracy and reliability of the magnitude estimates.

We also developed a dedicated duration magnitude (M_d) scale for northern Algeria, which marks the first such scale for the region based on a quite consistent amount of seismic data. Because, before, the CRAAG used a relationship that was developed for Southern California for duration magnitude estimation. Moreover, in this thesis we introduced a new tool for automatic duration measurement. The algorithm for estimating signal duration, based on the signal-to-noise ratio (SNR) within a moving window, provided stable and robust duration estimates.

The new M_d scale was derived through linear regression analysis, grouping data into various magnitude classes. Station correction coefficients were calculated for each site, and we compared the derived M_d values with M_L and M_w estimates. The results demonstrated excellent agreement between M_d and M_L , as well as a clear linear correlation between M_d and M_w , validating the effectiveness of using signal duration as an independent measure of earthquake size.

This work represents the first step in the implementation of automatic duration magnitude computation at the ADSN network. The methodology we developed can be easily integrated into the existing automatic procedures at ADSN for magnitudes calculation, with a potential for further refinement. Future improvements could include more rigorous criteria for record selection or the adoption of more robust statistical methods to enhance the accuracy and consistency of magnitude estimates.

Ultimately, the development of both the M_L and M_d scales for northern Algeria provides a

more reliable framework for earthquake magnitude estimation in the region. These new scales, well adapted to the region's specific seismic characteristics, will improve seismic monitoring and contribute to more accurate seismic hazard assessments. The findings of this work are crucial for enhancing the safety and preparedness of communities vulnerable to earthquakes in northern Algeria, while also supporting ongoing seismic research in the region. Furthermore, the methods developed here can serve as a model for other tectonically similar regions, offering insights into regional-scale magnitude estimation and risk mitigation.

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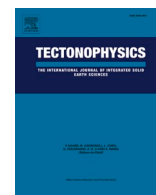
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Appendix A

Scientific contributions



A local magnitude scale (M_L) for Northern Algeria

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ABSTRACT

This study presents a local magnitude scale (M_L) based on the original Richter definition and designed for use within the Algerian Digital Seismic Network (ADSN). The magnitude scale is derived from the analysis of 17,377 zero-peak maximum amplitude traces extracted from the vertical component, simulated as Wood-Anderson seismograms. These traces are taken from a dataset of 1901 earthquakes recorded between January 1, 2010, and June 1, 2022, at a minimum of five stations in the ADSN network. To better account for the attenuation of direct and refracted waves in northern Algeria, amplitude decay analysis reveals the presence of two transition distances at 90 and 190 km, resulting in three segments. A distance correction term, $-\log_{10}(A_0)$, is introduced and described by the following trilinear function:

$$-\log(A_0) = \begin{cases} 0.6747 \cdot \log_{10}(R) + 0.0002 \cdot R + 1.6306 & R \leq 90 \\ 1.7736 \cdot \log_{10}(R) + 0.0002 \cdot R - 0.5169 & 90 < R \leq 190 \\ 2.4580 \cdot \log_{10}(R) + 0.0002 \cdot R - 2.0765 & R > 190 \end{cases}$$

R represents the hypocentral distance in kilometers. The derived distance correction formula provides a well-constrained M_L relationship for northern Algeria that is valid over a distance range of 5 to 600 km. Compared to other local magnitude relationships, the methodology proposed in this study consistently gives M_L values slightly higher than those calculated by the Southern California relationship over all distances, with an average difference of 0.2 units. We computed corrections for 72 stations by minimizing the M_L residuals. These corrections range from -0.50 to 0.54 , highlighting the influence of local site effects on the amplitude of the seismic signal. The magnitude residuals using our magnitude relationship and incorporating the station corrections, show that the standard deviation has improved significantly, from 0.34 to 0.24 . An M_L relationship specific to the northern Algerian region provides a valuable tool for seismic monitoring, hazard assessment, and earthquake research in the region.

1. Introduction

The magnitude is a fundamental seismological parameter that is closely linked to both earthquake shaking and seismic hazard. It is an integral part of numerous scientific activities, such as the creation of earthquake catalogs used, among other things, for the statistical analysis of earthquakes. These studies are essential for seismic hazard assessment and the calculation of parameters such as the Gutenberg-Richter b value (Wiemer and Wyss, 2002; Wyss, 2004; Ghosh et al., 2008). Since 1935 (Richter, 1935), a variety of magnitude scales have been developed.

Richter magnitude (also known as Local Magnitude, M_L) continues to be widely used as a significant reference point due to its simplicity and global reach, particularly in the media. M_L quantifies energy released by local earthquakes and evaluates seismic activity in specific regions (e.g., Richter, 1958; Kanamori and Jennings, 1978; Hutton and Boore, 1987; Boore, 1989). The local magnitude scale was originally created by measuring the highest horizontal amplitude, specifically from the zero to the peak of ground displacement on a Wood-Anderson seismogram. The Wood-Anderson seismometer is a horizontal torsion design with a natural period of 0.8 s, a magnification of 2800 , and a damping constant

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Unveiling complex fault geometry and driving mechanisms: insights from a refined data processing and multiplet analysis of the 2010 Beni-Ilmane seismic sequence (NE Algeria)

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SUMMARY

This paper offers a comprehensive re-analysis of the Beni-Ilmane 2010 seismic sequence, using a data set that is 100 per cent larger than previous studies. This unprecedented sequence in Algeria features three main shocks with magnitudes M_w 5.4, 5.1 and 5.1. Our approach involves high-precision relocation, which includes the development of a new 1-D minimum velocity model, followed by a double-difference (DD) procedure and hierarchical clustering. We determined the focal mechanisms (FMs) for 128 key events and identified 21 multiplet groups using an average cross-correlation threshold of 0.8. Our analysis offers new insights into fault geometry and addresses ongoing debates, by proposing a seismotectonic model that reveals the activation of 14 fault segments during the sequence, in contrast to previous oversimplified models that suggested two or three faults. The computed stress field from the inversion of 128 FMs aligns with a tectonic loading force due to the convergence of the African and Eurasian plates. These findings highlight the complexity of the fault network in the study area and shed light on the role of strike-slip faults in shaping the thrust belt. We found a strong link between multiplet groups and fluid movement along the fault network. Analysis of the temporal history of these multiplet groups provides new insights into fluid dynamics timescales, with an estimated hydraulic diffusivity (D) of $0.36 \text{ m}^2 \text{ s}^{-1}$ suggesting a fluid pressure diffusion process. The observed expansion of the aftershock area with the logarithm of time and the existence of repeating earthquakes indicates, for the first time, an aseismic slip mechanism that adds an additional layer to the driven processes. In conclusion, our results suggest that the underlying mechanisms governing the BI-2010 seismic sequence involve a complex interplay of tectonic loading, coseismic stress transfer, fluid dynamics and aseismic slip transients. We attempt to correlate our findings with various studies linking the structure, mechanics and fluid flow properties of fault zones and fault systems. The activation of smaller fault segments potentially averted a larger quake, resulting in three moderate main shocks and numerous aftershocks. This work not only enrich our understanding of seismic phenomena but also provides useful insights for seismic hazard assessment and risk mitigation strategies.

Key words: Fracture and flow; Earthquake dynamics; Induced seismicity; Dynamics and mechanics of faulting.

1 INTRODUCTION

The Beni-Ilmane earthquake sequence of May 2010, hereafter referred to as BI-2010, has been extensively studied for more than a decade. This sequence occurred on the southern front of the Tellian Atlas (Fig. 1a), within the transition zone between the Hodna

and Bibans mountains (Figs 1b and c), about 200 km from the plate boundary, as reported by Billi *et al.* (2023). The sequence, marked by three moderate main shocks (MS-1, MS-2 and MS-3 in Fig. 1b), caused damage to infrastructure and triggered subsequent aftershocks, inducing fear among the population for weeks. It began with MS-1, measuring M_w 5.4, on 14 May 2010, at 12:29 (GMT),

Time-dependent and spatio-temporal statistical analysis of seismicity: application on the complete data set of the 2010 Beni–Ilmane earthquake sequence

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SUMMARY

The Beni–Ilmane (BI) seismic sequence, situated in the north-central region of Algeria, began on 2010 May 14 with a main shock of M_L 5.4 followed by two other shocks on May 16 and 23 of M_L 5.3 for both. Using the complete data set from May 14 to August 31 and the Epidemic Type Aftershock Sequences (ETAS) model to separate background from aftershocks contribution gave a first insight around the uncertainty that surrounds the causes and mechanisms of the seismicity. According to our findings, two phases have been determined, the first one, beginning from May 14, shows low b - and p -values with most of the seismicity being self-triggered. In contrast, the second phase, beginning from May 23, demonstrates an increase of both the b - and p -values with a high number of background events. In the second phase, the background contribution involves 53 per cent of all seismic activity after May 23's peak which is high compared to typical main shock–aftershocks sequences. A possible explanation is that the main shocks began or assisted aseismic processes in addition to causing aftershocks. A sealed source at depth may have been broken by the third main shock, opening the way for the following incursion of highly pressurized fluids (hydrocarbures) confirmed by a high V_p/V_s ratios.

Key words: Africa; Spatial analysis; Earthquake hazards; Induced seismicity; Seismic tomography; Statistical seismology.

1 INTRODUCTION

The Tellian Atlas in northern Algeria has experienced numerous moderate to large earthquakes over the last two decades, beginning with the devastating earthquake in Boumerdes in 2003 May, which had a magnitude of M_w 6.8 (Yelles-Chaouche *et al.* 2004). These earthquakes are related to the oblique convergence between the African and Eurasian plates since the Late Cretaceous, with a rate of $4.5 \pm 0.3 \text{ mm yr}^{-1}$ along the N325 W $\pm 5^\circ$ shortening direction (Nocquet 2012, Fig. 1a). Indeed, the present strain partitioning in northeastern Algeria is the result of convergence involving three major fault structures, as indicated by the study conducted by Bougrine *et al.* (2019). First, the Offshore Fault System exhibits a slip deficit rate of 1.5 mm yr^{-1} . Secondly, the Gharimadou–North–Constantine Fault accommodates dextral displacement and has a significant present-day slip deficit rate estimated at 2.4 mm yr^{-1} . Lastly, a deformation rate of 1.5 mm yr^{-1} is observed along the fault system extending from the well-known E–W Gafsa Fault in

Tunisia to the South Atlas Fault System in Algeria, and further along the southwestern end of the NW–SE Aures Range until the Hodna Range near our study area (refer to Fig. 1b). Despite the relatively low deformation rate of 1.5 mm yr^{-1} and the main shock being considered a moderate event, the 2010 BI earthquake sequence generated a significant number of events during the first two weeks. Approximately 25 000 events were detected by a single nearby station, making it the most noteworthy seismic sequence in the northern part of Algeria.

One of the most significant features of this seismicity is the occurrence of seismic sequences with two or more main shocks and an important number of seismic events, which are mostly accompanied by spatial migration (Khelif *et al.* 2018; Yelles-Chaouche *et al.* 2018; Abacha & Yelles-Chaouche 2019; Bendjama *et al.* 2021; Boulahia *et al.* 2021; Abacha *et al.* 2022, 2023). The 2010 Beni–Ilmane (BI) earthquake sequence stands out as the most notable example due to the high number of seismic events. In Fig. 1(b), we present the recent main earthquake sequence that occurred in



Fluid-driven processes triggering the 2010 Beni-Illmane earthquake sequence (Algeria): evidence from local earthquake tomography and 4D Vp/Vs models

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Abstract The 2010 Beni-Illmane earthquake sequence, with 3 $M_w \geq 5.2$ mainshocks, occurred in the southern Tellian Atlas approximately 200 km southeast of the Algiers. We offer updated 3D Vp and Vs models, and we outline the space and time variations in the Vp/Vs ratio of the Beni-Illmane crust based on 2297 well-located aftershocks representing the complete aftershock sequence. Two main P-wave low-velocity patterns correspond to fault alignments determined from geological studies: the NE–SW thrust fault located at the foot of Djebel Choukhot and the NNW–SSE oriented Samma strike-slip

corridor. The former crosscuts the two P-wave high-velocity anomalies of the Kef Ouagnane and Kef Kherrat summits of the Choukhot Mountains. The third high-velocity anomaly represents the Jebel Affroun Mountain in the sub-Bibanic domain. Two high-Vp/Vs ratio patterns were revealed in the central aftershock part and along the NE–SW Ouanougha syncline, which constitutes a fluid reservoir and a fluid-saturated fault zone filled after aftershock activity. The 4D Vp/Vs models monitored fluid migration, and the modeled pore-pressure diffusion revealed that earthquake rupture along an ~N–S blind strike-slip fault fractured the closed reservoir and released high-pressure fluids, provoking a notable and long-lasting aftershock sequence. Our study underlines that local earthquake tomography (LET) can provide useful insights into the upward migration of fluids from a compartmentalized hydrocarbon reservoir.

Highlights

- This updated investigation used twice the number of events from a previous study.
- A high-Vp/Vs anomaly coincided with hydrocarbon reservoirs and fluid-saturated faults.
- The migration of high-pressure fluids released from a ruptured reservoir caused aftershock activity.

Supplementary Information The online version contains supplementary material available at <https://doi.org/10.1007/s10950-022-10130-8>.

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Keywords Beni-Illmane sequence · Seismotectonic · LET · Fluids · 4D Vp/Vs ratio, pore-pressure diffusion

1 Introduction

In May 2010, the Beni-Illmane region (north-central Algeria, Fig. 1a) was struck by a never-before-seen seismic sequence, as this was the first time in Algeria that a sequence was observed with three mainshocks of a $M_w \geq 5.2$ magnitude, recorded on May 14, 16,



The 24 January 2020 Mw 5.0 El Aouana Earthquake, Northeastern Algeria: Insights into a New NW–SE Right-Lateral Bejaia-Babors Shear Zone

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Abstract—On January 24, 2020, a Mw 5.0 earthquake occurred in the El Aouana region, northeastern Algeria. This region is located at the western end of the Lesser Kabylia Block (LKB), a rigid body that was weakly deformed during the late Cenozoic tectonic phase, and it is characterized by a lower seismic activity than that in its bounding regions. The mainshock focal mechanism was estimated via both the P-wave first motion and waveform modeling methods. The earthquake was associated with the rupture of a NW–SE-oriented right-lateral strike-slip fault, as revealed by a 6 km long and 2 km wide aftershock cluster. The seismic moment estimated through waveform modeling was 3.6×10^{16} Nm, while spectral analysis yielded a value of 3.9×10^{16} Nm corresponding to a magnitude of Mw 5.0, a source radius of 1.6 km, and a stress drop of 4 MPa. The spatiotemporal evolution of the aftershock sequence, as modeled using a restricted epidemic-type aftershock sequence (RETAS) stochastic model, yielded a slope $p = 1$, indicating that the earthquake was generated by tectonic forces and that the aftershock sequence included many subsequences. The calculated stress tensor suggested N–S compression, rotated clockwise relative to NW–SE Eurasia–Africa convergence. Finally, the recent seismic activity (2012–2021) and geological observations in the area led to the suggestion of a new NW–SE right-lateral shear zone, namely, the Bejaia-Babors shear zone, which was incorporated into a seismotectonic growth model involving slip along inherited E–W structures. The pattern of stepover structures throughout this wide shear zone was enhanced during the recent seismic evolution.

Keywords: El Aouana, RETAS, source parameters, stress tensor, shear zone, stepovers.

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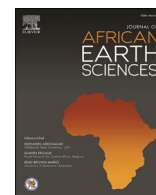
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1. Introduction

Northeastern Algeria is subjected to substantial, moderate, and shallow seismic activity due to its location along the African–Eurasian plate boundary (blue line in Fig. 1a). One of the main statistical characteristics of Algerian seismicity is its concentration in the country's eastern part, where nearly two-thirds of the recorded seismic events have occurred (Abacha, 2015; Yelles-Chaouche et al., 2006). In this territory, numerous seismic sequences took place over the past decade (2012–2021), particularly around the Lesser Kabylia Block (LKB) (Fig. 1b). These include the 2017 seismic sequences along the Mcid Aïcha-Debbagh Fault (MADF) in the south (Bendjama et al., 2021), the 2014 Ziamia and 2019 Jijel earthquakes in the north (Yelles-Chaouche et al., 2021), and the 2012–2013 Bejaia-Babors sequences in the west (Boulahia et al., 2021).

On January 24, 2020, at 07 h 24 m 20 s, the region of Jijel was shaken once more by a shallow earthquake of magnitude Mw 5.0, namely, the El Aouana earthquake at the western boundary of the LKB, 20 km southwest of Jijel (Fig. 1c). The western transverse fault system of Bejaia-Babors constitutes a NW–SE-trending structure associated with right-lateral strike-slip earthquake focal mechanisms. Recent advances in understanding the fault geometry mechanism in this region have been obtained by Boulahia et al. (2021), who identified the majority of the currently known en echelon fault segments that define the western boundary of the transverse fault system. The El Aouana 2020 earthquake exhibits a focal mechanism with one of the planes subparallel to the eastern boundary of the transverse fault system.



The 13 July 2019 Mw 5.0 Jijel Earthquake, northern Algeria: An indicator of active deformation along the eastern Algerian margin

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ABSTRACT

On July 13, 2019, a moderate earthquake shook the Jijel region of Algeria, ~300 km east of Algiers (Capital of Algeria). In the past, the Jijel region experienced a destructive earthquake (I_0 : X) on 21–22 August 1856, triggering a moderate tsunami in the western Mediterranean region. The 2019 event is the second important event to have occurred along this part of the Algerian margin, which has been undergoing an inversion process since the Pliocene Epoch. The event occurred 40 km to the north of Jijel City, at 10 km depth. No property damage or injuries were associated with this event. The seismic moment release estimated from waveform modeling was $M_0 = 3 \times 10^{16}$ Nm, corresponding to a Mw 5.0 event. The mainshock focal mechanism is associated with the rupture of an E–W thrust fault that is clarified by a 4-km-long $\times$ 10-km-wide aftershock cluster along the continental slope of the Jijel margin. In the same area, ~30 km to the west and offshore of the coastal village of Zia-ma–Mansouriah (west of Jijel) a Mw 4.1 earthquake occurred in 2014. This event is also associated with the rupture of a western contiguous E–W thrust fault. Stress tensor inversion produces a main (σ_1) maximum stress axis oriented N356E, which is in agreement with previous stress studies of the Lesser Kabylia–Babors block. The present-day stress field is characterized by a compressional tectonic regime ($R' = 2.96 \pm 0.21$). Coulomb stress analysis of the 2014 Zia-ma and 2019 Jijel earthquakes indicates that the two earthquakes appear to be individually triggered events with no stress transfer between them. The 2019 Jijel earthquake is associated with one of the three active *en echelon* faults that have already been identified in the central part of the Algerian margin via recent marine surveys. It belongs to the serie of many moderate seismic events that have recently affected the Algerian margin (Boumerdes, Algiers, Bejaia), and is representative of the active deformation process along the Algerian margin that is marked by the initiation of an incipient subduction zone with a progressive plate boundary migration since the late Miocene.

1. Introduction

The Algerian margin seismicity has been poorly constrained for decades due to the lack of a high-performance seismic network and marine investigations. Few studies have investigated the deformation process that affects this region crosscutting by the African–Eurasian plate boundary of the western Mediterranean (Fig. 1a).

The offshore region of northern Algeria was the site of a number of historic marine seismic events, such as the 1365 Algiers Earthquake (Mokrane et al., 1994) and 1790 Oran Earthquake (Benhallou, 1985; Chimouni et al., 2018). However, archives and historical reports have provided only sparse descriptions of these events. The 1856 Jijel

earthquake sequence on 21–22 August remains one of the best known historical offshore events (Benhallou, 1985; Yelles-Chaouche et al., 2009b; Harbi et al., 2003). A moderate tsunami was generated from one of the mainshocks of this sequence, and impacted the Mediterranean coast. The recent 1989 offshore Ms 6.0 Tipaza Earthquake (Bounif et al., 2003) and 2003 Mw 6.8 Boumerdes Earthquake (Yelles-Chaouche et al., 2004) occurrence demonstrated the crucial importance for more marine investigations along the Algerian margin (Fig. 1b) to better understand the deformation pattern along the African–Eurasian plate boundary in the western Mediterranean region.

To address the urgent need to improve our knowledge of Algerian margin geodynamics, several marine projects have recently

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Exploring the Structural Components of the Bejaia-Babors Shear Zone (BBSZ) in NE Algeria: Evidence from Local Earthquake Tomography using Recent Seismic Events (2012-2022).

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Abstract:	In recent years, the Algerian region of Bejaia-Babors (BB) has experienced significant seismic activity, including the Bejaia-Babors seismic sequence in 2012-2013, the Jijel earthquake in 2019, the El Aouana earthquake in 2020, and Bejaia subsequent earthquakes in 2021 and 2022. These seismic events have not only brought to light the existence of the Bejaia-Babors Shear Zone (BBSZ) but have also emphasized the importance of discerning its structural components, depth, and extent. Our study focuses on the analysis of seismological data from 2012 to 2022, with a particular emphasis on elucidating the intricacies of this geological structure. Using the LOTOS (local tomography software) algorithm, we conducted three iterations of tomographic inversion, successfully obtaining horizontal and vertical sections that facilitated the identification and characterization of subsurface anomalies. The resulting 3D velocity models unveiled key tectonic structures within the BBSZ, including the Offshore Faults System of Jijel (OFSJ), South Greater Kabylia Fault (SGKF), Transversal Fault 1 (TF1), and the collision between the Lesser Kabylia Block (LKB) and the Babors (THF-1). Furthermore, brittle-ductile shears were identified along the Aftis Fault (AF) in the east and brittle shears along the Babors Transverse Fault (BTF) in the west. P-wave velocity analysis indicated the presence of rigid blocks. The observed high V_p/V_s ratio near segment 3 of the BTF fault suggests the presence of a fluid reservoir, likely involved in the Bejaia-Babors seismic sequence (2012-2013), as previously documented. These findings provide valuable insights into the tectonic framework of the BBSZ, highlighting major fault systems and the interaction between different tectonic blocks. The presence of brittle-ductile shears along the AF suggests complex deformation processes in this region. Overall, by identifying key fault systems, characterizing subsurface anomalies, and unveiling the presence of fluid reservoirs, our research not only contributes significantly to geodynamic knowledge but also holds immense significance for seismic hazard assessment, resource exploration, and future research in this field.
Suggested Reviewers:	Haakon Fossen, Prof haakon.fossen@uib.no Hakon Fossen is a prominent structural geologist who has conducted extensive research on shear zones Ivan Koulakov, Prof

Natural Hazards

Detecting and Interpreting Earthquake Swarms in Northeastern Algeria (2007-2020): Evidence from a Dual Approach. --Manuscript Draft--

Manuscript Number:	NHAZ-D-24-00416
Full Title:	Detecting and Interpreting Earthquake Swarms in Northeastern Algeria (2007-2020): Evidence from a Dual Approach.
Article Type:	Manuscript
Keywords:	Northeastern Algeria, Earthquake swarms, ETAS, Fluid migration, Moment release
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Abstract:	In northeastern Algeria, seismic activity is distinguished by the occurrence of swarms or classic sequences, showing intense temporal and spatial clustering, lasting from days to months and extending over distances of a few to tens of kilometers, sometimes displaying migration patterns over several weeks. Categorizing seismic swarms and comprehending their underlying mechanisms is essential for mitigating seismic risk in the region, especially as they are indicative of potential physical changes in the crust, such as slow slip or fluid flow, which makes them particularly noteworthy. To identify seismic swarms, two methodologies were applied, including the use of spatio-temporal ETAS models and moment release time series analysis, covering data from 2007 to 2020 in the region. Identifying 32 clusters, totaling 676 events and accounting for 29% of the catalog above the magnitude threshold of 2.8, revealed 32 swarm sequences marked by event counts exceeding five, with subsequent computation of the b-value across the entirety of the swarm catalog; these clusters were then classified into zones based on the geological and tectonic context of each area. Ultimately, we offer potential explanations for the mechanisms driving the formation of these swarms, alongside investigating their spatial migration and temporal evolution.
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Acta Geophysica

Location and Magnitude from Seismic Intensity Data of Historical earthquakes in the North-eastern part of Algeria --Manuscript Draft--

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Full Title:	Location and Magnitude from Seismic Intensity Data of Historical earthquakes in the North-eastern part of Algeria
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Abstract:	We determine the locations and intensity magnitudes MSI for a test-set of 10 earthquakes that occurred between 1856 and 1980, by adapting and calibrating the method proposed by (Backun and Wentworth, 1997). We analyzed a total of 1082 intensity observations for 14 “training-set” recorded and located events, from the period between 1949 and 2003, to determine the intensity magnitude MSI corresponding to the surface wave magnitudes Ms. The intensity magnitude MSI is derived from the mean of the attenuation relation $MSI = (EMS + 0.6445 + (0.0271 * \Delta i)) / 1.3468$, where Δi is the distance in kilometer of the EMS (European Macroseismic Scale) observations. The epicentral region of each event is bounded by contours of rms [MSI] with 50%, 67%, 80%, 90%, and 95% confidence-level, for location, on a 51 x 51 trial epicentres grid with 5km grid spacing. To validate our data we tested also the software Boxer 42.1 implemented by (Gasperini et al., 2010) and based on the methods of (Pasolini et al., 2008) to estimate epicentre location and magnitude. The obtained results demonstrate the effective applicability of both methods on the Algerian Macroseismic Database of historical earthquakes.
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