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THERMOTECTONIC HISTORY OF THE SOUTHERNMOST NORTHERN ANDES

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DEDICATION

To my family, Mayito, and my friends: José, Peña, Andrés, Sebas, Yei, La Flaca, Isabelingo, Fer, Alice, Eli, Liying, Xiaoqin, Riccardo, Patrick, Dan, Fatime, Jorge, Felipe, Silvia and Mayank; thank you for your unconditional support and for making everything a little bit easier along the way. I can be called a Doctor now, and it's because of you.

ABSTRACT

The southernmost segment of the Andes, which encompasses southern Ecuador and northern Peru, is a geologically critical part of the Andean belt. This region marks a significant change in the mountain chain's orientation from northeast to northwest and lies adjacent to areas in the north that have experienced substantial oceanic terrane accretion. Consequently, the region has undergone prolonged and complex geodynamic evolution from the Paleozoic to the Cenozoic, making it a pivotal area for comprehending mountain-building processes and magmatism at convergent plate boundaries.

This study uses U-Pb geochronology of detrital and bedrock zircon, as well as whole-rock and zircon geochemistry, Hf and O isotope analyses, apatite fission track dating, and numerical thermo-tectonic modeling with Pecube and 4DTherm to elucidate the thermo-tectonic evolution of this area.

Provenance analysis reveals that the Neoproterozoic to Triassic history of the region aligns more closely with northern Gondwanan signatures, such as the Brasiliano/Pan-African orogenies and the opening of the Rheic Ocean, than with classic Andean signals, like the Pampean and Famatinian orogenies. These findings suggest that, during this period, the area experienced crustal reworking processes with limited mantle input. Therefore, the region functioned more as a zone of crustal refinement than of new crust formation.

The results also indicate that the region experienced a predominantly extensional regime during the Mesozoic era, accompanied by an increase in mantle-derived contributions to magmas and a decrease in crustal signatures. By ~110 Ma, slab rollback produced the migration of the arc to the west, which led to the development of an oceanic-type basement during the middle to late Cretaceous, which was coeval with the opening of the Celica-Lancones Basin.

The Cenozoic evolution is characterized by sustained magmatic activity between ~60 and 20 Ma and a gradual shift in magma sources. Two-dimensional thermal history modeling reveals fault-controlled exhumation in southern Ecuador with three distinct cooling episodes: Middle Eocene, Late Eocene–Early Oligocene, and Late Oligocene–Early Miocene, varying across different crustal domains. Coeval magmatic and exhumation-related cooling is also evident. 4DTherm modeling indicates plutons with overlapping U-Pb and AFT ages were emplaced at ~3000–3700 m, showing a stark contrast between rapid post-magmatic (577–613 °C/Ma) and slower exhumation-related cooling (3.6–7.4 °C/Ma).

SOMMARIO

Il segmento più meridionale delle Ande, che comprende l'Ecuador meridionale e il Perù settentrionale, è una parte geologicamente critica della cintura andina. Questa regione segna un cambiamento significativo nell'orientamento della catena montuosa da nord-est a nord-ovest e si trova adiacente ad aree del nord che hanno sperimentato una sostanziale accrezione di terrane oceaniche. Di conseguenza, la regione ha subito una prolungata e complessa evoluzione geodinamica dal Paleozoico al Cenozoico, che la rende un'area cruciale per la comprensione dei processi di costruzione delle montagne e del magmatismo ai confini convergenti delle placche.

Questo studio utilizza la geocronologia U-Pb di zirconi detritici e bedrock, nonché la geochemica di rocce intere e zirconi, le analisi isotopiche di Hf e O, la datazione della traccia di fissione dell'apatite e la modellazione numerica termotettonica con Pecube e 4DTherm per chiarire l'evoluzione termotettonica di quest'area.

L'analisi della provenienza rivela che la storia dal Neoproterozoico al Triassico della regione si allinea più strettamente con le firme del Gondwan settentrionale, come le orogenesi brasiliana/pana-africana e l'apertura dell'Oceano Retico, che con i classici segnali andini, come le orogenesi pampeana e faminiana. Questi risultati suggeriscono che, durante questo periodo, l'area ha subito processi di rielaborazione crostale con un limitato apporto di mantello. Pertanto, la regione ha funzionato più come zona di raffinamento crostale che di formazione di nuova crosta.

I risultati indicano inoltre che la regione ha sperimentato un regime prevalentemente estensionale durante il Mesozoico, accompagnato da un aumento dei contributi di origine mantellare ai magmi e da una diminuzione delle firme crostali. Entro ~110 Ma, il rollback dello slab ha prodotto la migrazione dell'arco verso ovest, che ha portato allo sviluppo di un basamento di tipo oceanico durante il Cretaceo medio-tardivo, coevo all'apertura del bacino Celica-Lancones.

L'evoluzione cenozoica è caratterizzata da una sostenuta attività magmatica tra ~60 e 20 Ma e da un graduale spostamento delle fonti magmatiche. La modellazione bidimensionale della storia termica rivela un'esumazione controllata da faglie nell'Ecuador meridionale con tre distinti episodi di raffreddamento: Eocene medio, tardo Eocene-inizio Oligocene e tardo Oligocene-inizio Miocene, che variano in diversi domini crostali. È evidente anche un raffreddamento magmatico coevo e legato all'esumazione. La modellazione 4DTherm indica che i plutoni con età U-Pb e AFT sovrapposte sono stati depositati a ~3000-3700 m, mostrando un netto contrasto tra il rapido raffreddamento post-magmatico (577-613 °C/Ma) e quello più lento legato all'esumazione (3.6-7.4 °C/Ma).

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CHAPTER 1: INTRODUCTION

Subduction settings are areas on Earth where a plate dives below another tectonic plate (oceanic or continental) into the mantle, and are considered to be one of the most complex geological environments on our planet (Lallemand & Heuret, 2017). Most subduction zones are characterized by four common features: a trench, a slab, and a magmatic arc (Schellart, 2023).

The development and evolution of subduction zones are influenced by both continental-scale and localized processes, creating complex and heterogeneous orogens. A prominent example of this is the Andean belt, which formed along the western margin of South America, extending from Colombia in the north to southern Chile (Fig. 1.1). The Andes are often described as a fragmented orogen due to significant variations along strike, leading to their classification into northern, central, and southern segments (i.e., MéGard, 1987; Ramos, 2010; Ramos & Aleman, 2000; Val & Willenbring, 2022).



Figure 1.1 Map of western South America illustrating the subduction plate boundary, oceanic aseismic ridges, flat-slab subduction zones, segments of the Andean Mountain chain, and the study area (modified from Horton, 2018).

These segments are further classified into two primary types based on the composition of their underlying basement (MéGard, 1987). The northern segment, referred to as the "Cordilleran" Andes, is characterized by oceanic rocks primarily derived from accreted terranes (Cediel, 2019; MéGard, 1987; Vallejo et al., 2009). In contrast, the central and southern segments, known as the "Liminal or Marginal" Andes, are predominantly composed of magmatic rocks emplaced on continental crust (Aubouin, 1972; Mégard, 1984). Each of these segments has experienced a distinct, long, and varied tectonic history shaped by the interplay of geological processes such as terrane accretion, subduction dynamics, and episodic magmatism (Ramos, 2010, 2018). These processes have given rise to diverse geological features and tectonic structures across the region, making the Andes a focal point for extensive geological research (Faccenna et al., 2017; Horton & Folguera, 2022; Oncken et al., 2006; Ramos, 2009). Nevertheless, certain areas remain poorly studied, particularly the southernmost portion of the northern Andes, such as southern Ecuador and northern Peru, as highlighted by several recent studies (Martínez-Ardila et al., 2023; Siravo et al., 2021; Spikings et al., 2015; Witt et al., 2023; Witt et al., 2019). This region is particularly interesting as it marks a significant change in the strike orientation of the orogen, commonly known as the Huancabamba Deflection (Gansser, 1973), where it shifts from N20E in the north to N20W in the south (Mourier, 1988).

The main objective of this dissertation is to provide new insights into the tectonic evolution of the southernmost northern Andes. First, this study aims to reconstruct the tectonic background of the area, including its Neoproterozoic and Paleozoic history, through provenance analysis of detrital zircons from forearc basin sediments. It then seeks to unravel the Mesozoic and Cenozoic history by studying the magmatic arc, analyzing its age, spatial distribution, geochemistry, and cooling history of in-situ granitoids. The methodologies employed in this research include U-Pb dating in zircon, whole-rock geochemistry, zircon geochemical analysis, zircon isotopic analysis (EHf_t and $\delta^{18}\text{O}$), fission-track dating in apatite, and (U-Th)/He dating in apatite, and numerical modelling using Pecube (Braun, 2003).

1.1 Morpho-structural configuration of the southernmost northern Andes

The South Ecuadorian–North Peruvian region displays significant morpho-structural diversity, as it comprises a variety of distinct geological domains (Fig. 1.2), which extend from west to east into: i) the coastal lowlands, which are characterized by Cretaceous-Tertiary forearc basins, and the coastal range represented by the Amotape-Tahín block and the Peruvian Coastal batholith; ii) the Western Cordillera, overlain by Cenozoic sediments and volcanics; iii) the hinterland regions, consisting of a central depression filled with Late Miocene to Recent volcano sedimentary rocks; iv) the Eastern

Cordillera, composed of folded and faulted sediments, which is also known as the Cordillera Real in Ecuador; and v) the Andean Foreland, which can be subdivided into west-dipping thrust slices and large antiforms of the Sub-Andean zone to the west (Litherland, 1994), and the folded Meso-Cenozoic sediments from the Oriente Basin in Ecuador, the Marañón and Huallaga Basins in Peru, to the east (Jaillard, 2022).

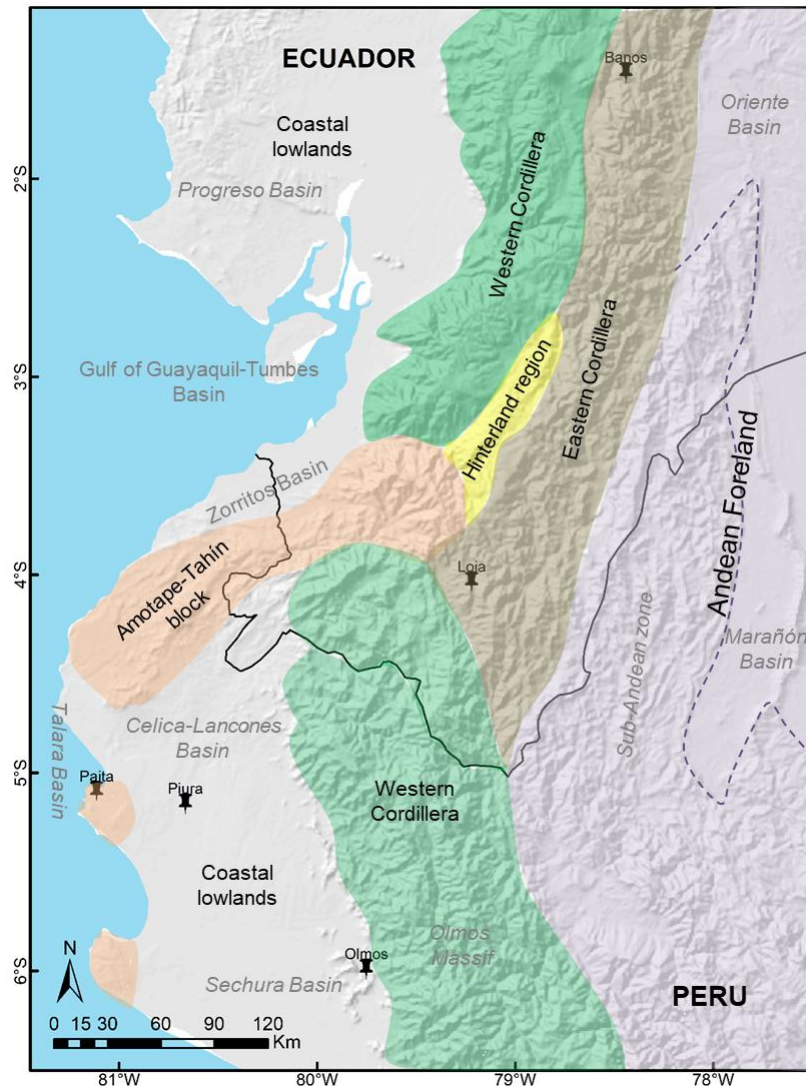


Figure 1.2 Morpho-structural domains of southwestern Ecuador and northwestern Peru.

1.1.1 Forearc basins

The forearc region includes both the submerged Andean slope and the emergent coast, bounded by the coastal batholith and the western cordillera to the east and the Peru-Chile trench to the west (Jaillard, 2022; and references therein). Forearc basins, which crop out along the Pacific coast and in the Pacific lowlands are filled with Tertiary marine and continental sedimentary deposits. From north to south, these basins include the Ecuadorian Progreso and Gulf of Guayaquil-Tumbes basins, as well as, the Peruvian Zorritos, Talara, Celica-Lancones and Sechura basins (Fig. 1.2).

1.1.1.1 Ecuadorian forearc basins

The forearc region of southern Ecuador comprises the Progreso and Gulf of Guayaquil-Tumbes basins, which developed on oceanic terranes accreted to the South American continent during the Late Cretaceous (Aizprua et al., 2020; Jaillard, 2022; Vallejo et al., 2009).

The Progreso Basin is an Oligocene to Recent forearc basin with a distinct triangular shape (Bulois et al., 2023; and references therein). It is filled with thick siliciclastic sequences that exhibit a wide range of facies, from alluvial fans and fan deltas to inner and outer shelf environments, all of which have been extensively reworked at the shoreline by waves and currents (Alemán et al., 2021).

South of the Progreso Basin lies the Plio-Pleistocene Gulf of Guayaquil-Tumbes Basin, which exhibits a complex architecture, consisting of three main depocenters: the Esperanza, Jambelí and Tumbes sub-basins (Egbue & Kellogg, 2010). These sub-basins are separated by structural highs, which are likely the result of transpression associated with strain partitioning along a major zone of weakness (Witt & Bourgois, 2010). The evolution of the basin is closely linked to extensional and strike-slip processes driven by the tectonic escape of the North Andean Block (or North Andean Sliver), a crustal fragment accreted onto the continent since the Late Cretaceous (Jaillard et al., 1999b).

1.1.1.2 Peruvian forearc basins

The northern Peruvian foreland includes the Zorritos, Talara, Celica-Lancones and Sechura basins. The Zorritos Basin, consists of an Upper Tertiary sedimentary sequence of shales, sandstones and conglomerates. The southern margin of this basin exposes a predominantly Miocene sedimentary sequence that dips gently to the NNW (Witt & Bourgois, 2010). The history of the basin is linked to regional tectonic features, including the Pilar de Zorritos geoform, a southwest-northeast trending structural high associated with the Dolores-Guayaquil Megashear (Fildani et al., 2008).

The Talara Basin, comprises approximately 9 km of Upper Cretaceous to Eocene units deposited in environments ranging from deep water to fluvial (Fildani et al., 2008). This basin records deformation resulting in the formation of fault-bounded blocks with significant thickness variations (Lajo-Yáñez et al., 2022). The basin encompasses both onshore and offshore regions, with the Amotape-Tahuin Massif serving as a prominent feature separating the offshore Talara Basin from the inland Celica-Lancones Basin.

The Celica-Lancones Basin is filled by Early Cretaceous to Eocene marine volcanic and sedimentary sediments with a siliciclastic composition to the west and volcanoclastic to the east (Jaillard et al.,

1999a). This basin is thought to have formed during the Early Cretaceous in an extensional setting driven by roll-back processes of the subducting Farallon Plate (Winter, 2008).

The Sechura Basin contains Late Eocene to Pleistocene sedimentary units. These units were deposited during major transgressive cycles (Dumbar et al., 1990) and exhibit a complex sedimentary history, with depositional environments transitioning from marine-influenced lagoons to arid conditions (Christol et al., 2017).

1.1.2 Coastal range

1.1.2.1 The Amotape-Tahuin block

The Amotape–Tahuin Massif is a tectonostratigraphic block that trends east–west in Ecuador and shifts to a north–south orientation in Peru. This is composed of Palaeozoic metasedimentary rocks, Triassic plutonic rocks and Jurassic high-pressure metamorphic rocks (Noble et al., 1997; Riel et al., 2014; Spikings et al., 2005). The high-pressure metamorphosed oceanic rocks are interpreted as exhumed subduction complex or accretionary prism deposits, with muscovite $^{40}\text{Ar}/^{39}\text{Ar}$ cooling ages of about 129 Ma (Arculus et al., 1999; Litherland, 1994; Spikings et al., 2015).

The origin and structural evolution of the massif remains controversial. Mourier (1988) proposed that it is an allochthonous continental terrane accreted during the Early Cretaceous. Later, Ramos (2009) proposed an earlier accretion to the South American margin during the Permian. More recent studies have hypothesized that the Amotape-Tahuin Massif separated from the South American margin during the opening of the Celica-Lancones basin and subsequently rotated 90° clockwise (Espurt et al., 2018; Riel et al., 2014; Witt et al., 2017).

1.1.2.2 The Peruvian Coastal Batholith

The Peruvian Coastal Batholith consists of over 1000 plutons, predominantly granodiorite and tonalite, with ages ranging from 100 to 34 Ma (Mukasa, 1986; Pitcher, 1985). The majority of its emplacement occurred during the mid- to late Cretaceous (Giletti & Day, 1968). This batholith forms a linear belt of calc-alkaline granitoids. Cobbing and Pitcher (1983) originally divided it into five distinct segments based on compositional variations: Toquepala (18–16.5°S), Arequipa (16.5–12°S), Lima (12–9°S), Trujillo (9–6.5°S), and Piura (6.5–4°S). More recently, Ccallo Morocco et al. (2021) proposed reducing these five segments to three: Piura (4–6°S), Lima (6–14.5°S), and Arequipa (14.5–18°S), with two transition zones between segments: Trujillo-Chiclayo (6–8.5°S) and Chaparra-Caraveli (14.5–16°S).

1.1.3 The Western Cordillera

The Ecuadorian Western Cordillera exhibits a complex geological history shaped by tectonic processes, including accretion, volcanic activity, and sedimentation (Vallejo et al., 2020; and references therein). Its composition is primarily dominated by Early to Late Cretaceous oceanic plateau basalts, ultramafic rocks, and marine turbidites (Aspen & Litherland, 1992). This prominent high-relief feature also contains an early Eocene basaltic-to-andesitic oceanic island arc, a Paleocene-Eocene marine turbidite basin fill, and a late Eocene-Oligocene terrestrial sequence, all intruded by mid-Eocene and younger I-type granitoids (Hughes & Pilatasig, 2002).

In contrast, the Peruvian Western Cordillera consists of Mesozoic to Tertiary rocks with no direct evidence of oceanic accretion (Wipf, 2006). In northern Peru, it is characterized by the steeply folded and faulted terrain of the Olmos massif. Lithologically, it is composed of Neoproterozoic shales and schists of the Olmos Complex and Salas Formation, limestones of the La Leche and Sávila Formations, and quartzites of the Neocomian Goyllarisquizga Formation (Jaimes Salcedo, 2011).

1.2 General stratigraphy of the southernmost northern Andes

The stratigraphy of the region reflects a complex geological history shaped by transgressions, regressions, changes in subduction dynamics, and terrane accretion. To provide a comprehensive overview, Fig. 1.3A illustrates the general stratigraphic column of southwestern Ecuador, while Fig. 1.3B summarizes the stratigraphic column of northwestern Peru.

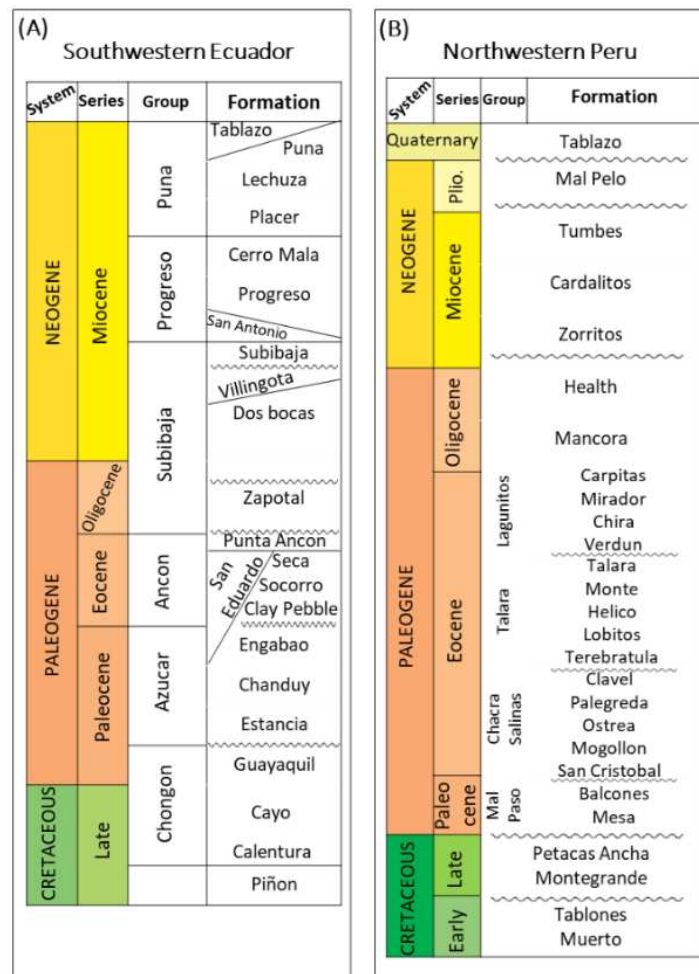


Figure 1.3 Summarized stratigraphic columns. (A) Southwestern Ecuador, modified from Ordoñez et al. (2006). (B) Northwestern Peru, modified from Kraemer et al. (1999). Wavy lines indicate periods of erosion or non-deposition.

1.3 Overview of the Late Mesozoic- Cenozoic geodynamic background

The northern Andes experienced alternating phases of extension and contraction, resulting in the formation of deep half-graben basins during extensional periods, and the subsequent inversion of normal faults and development of regional unconformities during compressional phases. Although these events have been described by various authors, the original nomenclature established by Steinmann (1929) has frequently been used or modified. In the following paragraphs, I aimed to summarize the primary tectonic events and, where feasible, draw connections between the broader "Andean events" with more localized tectonic events specific to the southernmost northern Andes.

The late Mesozoic is marked by the Late Cretaceous "Peruvian" tectonic phase, a widespread deformation event that affected the entire western margin of South America (Jaillard & Soler, 1996; Mégard, 1984). In the northern Andes, this event (~75–80 Ma) stands out as a major tectonic episode. Due to limited age constraints, early works by Steinmann (1929) and Mégard (1984) did not distinguish between the Santonian and Late Campanian events, merging them into the "Peruvian tectonic phase". However, recent studies have shown that, in the northern Andes, the Peruvian phase

is a distinct tectonic event characterized by the accretion of the Caribbean oceanic plateau along the continental margin during the Late Campanian (75–72 Ma; Jaillard et al., 2009; Luzieux et al., 2006; Vallejo et al., 2009). While there is general consensus regarding the initial time of accretion and the age of the plateau, the nature of the accretion process remains debated. Some propose a single accretion phase occurring between 75–65 Ma (Vallejo et al., 2019; Vallejo et al., 2009), whereas others suggest multiple episodes of accretion at ~75 Ma, ~68 Ma, and ~58 Ma (Hughes & Pilatasig, 2002; Jaillard et al., 2009; Lebras et al., 1987).

Subsequently, the Paleocene–Early Eocene 'Incaic phase' was a major compressional event, often regarded as the primary shortening episode in the Peruvian Andes. It produced significant deformation across the Western and Eastern Cordilleras, leading to the development of the Marañón fold-and-thrust belt in north-central Peru (McKee & Noble, 1990; Noble et al., 1990; Quang et al., 2005), and the exhumation of Late Cretaceous to early Paleogene plutons (Noble et al., 1979). This phase has been subdivided into two discrete episodes: Incaic I and II (Bussell, 1983; McKee & Noble, 1990). The poorly constrained Incaic I phase (~60 Ma), likely resulted from a shift in the convergence direction of the South American and Farallon plates (Noble et al., 1990). In contrast, the Incaic II phase, dated to the Paleocene or early Eocene (Noble et al., 1979), has been linked to a reorganization of Pacific plate motion, recorded by the Hawaiian-Emperor elbow (Gordon & Jurdy, 1986).

The period between the Late Eocene and Late Oligocene has been interpreted as a phase of slab rollback and extensional tectonics (45–30 Ma; Witt et al., 2023; and references therein). In contrast, northwestern Peru experienced a major compressional event during the same period, marked by uplift and a shift from marine to fluvial-deltaic depositional environments (36–27 Ma; Espurt et al., 2018; Naeser et al., 1991; Noble et al., 1990). This tectonic episode culminated in the Aymara event (28–26 Ma), which is linked to the breakup of the Farallon Plate into the Cocos and Nazca plates around 23 Ma (Lonsdale, 2005; Meschede & Barckhausen, 2001).

The Miocene-Pliocene "Quechua phases" (I, II, and III) were events of extensive crustal thickening and uplift throughout the Andes (Mégard, 1984; Noble et al., 1990). The "Quechua I" phase is loosely constrained between 21.5 and 12.5 Ma, in central Peru (Jaillard et al., 2000; McKee & Noble, 1990). However, this compressional period appears to have been shorter in the Andes of Ecuador and Peru, as evidenced by the opening of Early to Middle Miocene extensional intermontane basins (16–14 Ma; Jaillard et al., 2000; and references therein). In Ecuador, these basins were filled with Middle Miocene fine-grained lacustrine deposits (14–10 Ma). In northern Peru, this interval is particularly significant due to the approach and subduction of the "Lost Inca Plateau," which reached the trench at approximately 13–10 Ma (Gutscher et al., 2000; Rosenbaum et al., 2005).

The "Quechua II" phase lasted from about 10 to 8 Ma (Mégard, 1984; Sébrier & Soler, 1991) and marked the onset of eastward thrusting of the Paleo-Andes onto the Brazilian and Guiana shields, initiating active deformation in the Sub-Andean thrust and fold belts (Baby et al., 2025; Jaillard et al., 2000). In Ecuador, this phase is characterized by the compressional closure of the intermontane basins (Steinmann et al., 1999), signaling a transition from an extensional to a compressional regime around 9-8 Ma. Although timing constraints are less precise in northern Peru, the shift to a compressional regime is also thought to have occurred in the Late Miocene (Jaillard et al., 2000).

Finally, the contractional "Quechua III" phase occurred between 7 and 4 Ma, driven by east-west shortening across the entire Andean chain, resulting in the development of the Subandean fold-and-thrust belt. In Ecuador, deformation and uplift during this phase were particularly influenced by the subduction of the Carnegie aseismic ridge. The timing of the ridge's arrival in the subduction zone remains debated, with estimates ranging from 2 Ma (Lonsdale, 1978), 5–6 Ma (Collot et al., 2022; Margirier et al., 2022), 2–8 Ma (Gutscher et al., 1999), to as early as 15 Ma (Margirier et al., 2023; Spikings et al., 2010).

1.4 Thesis outline

Chapter 2 provides a concise overview of the sample collection procedures and preparation techniques for subsequent analyses. It also details the analytical methodologies employed, including the specific conditions, instruments, and laboratories used. The results obtained from each method are summarized and form the foundation for the interpretations presented in the following chapters.

Chapter 3 explores the Neoproterozoic to Paleozoic history of southern Ecuador and northern Peru through provenance analysis of detrital zircons collected from the Manabí, Progreso, and Gulf of Guayaquil–Tumbes basins in Ecuador, and the Talara and Zorritos basins in Peru. It examines the geochemical and isotopic signatures of these grains and situates the results within a broader tectonic framework, contributing to a more comprehensive understanding of the geological evolution of the region.

Chapter 4 outlines the tectonic evolution of the Mesozoic magmatic arc in northern Peru, focusing on its spatial and temporal distribution, the magma sources responsible for the Mesozoic intrusions and the tectonic regimes that influenced the region during this period.

Chapter 5 explores the geochemical and petrogenetic characteristics of some Cenozoic plutons in northern Peru. It also presents the first U-Pb ages for these plutons, along with new zircon isotopic data, including ϵHf_t and $\delta^{18}\text{O}$ values. These datasets are used to infer the nature of the magmatic reservoirs that may have contributed to their formation.

Chapter 6 reconstructs the exhumation history of southern Ecuador using apatite fission-track thermochronology. It examines distinct structural and geological domains to assess whether exhumation followed a migratory pattern, better constrain the timing of cooling events, quantify cooling rates through time, and identify the main driving mechanisms behind them.

Chapter 7 outlines the numerical modeling approaches used in this study, including: (i) Pecube, to reconstruct the thermal and exhumation history of one fault-bounded region in southern Ecuador—the Zabanilla Unit and Zamora Batholith in the Real Cordillera; and (ii) 4DTherm modeling, applied to distinguish between the magmatic and exhumation histories of dated intrusions within the Ecuadorian Cenozoic arc.

Finally, Chapter 8 presents the concluding remarks, summarizing the key findings of the project, highlighting its main contributions, and proposing directions for future research.

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CHAPTER 2: METHODOLOGY & RESULTS

2.1 Sample collection

For this project, we collected 41 granitoid samples and 5 metamorphic bedrock samples from southern Ecuador and northern Peru. Additionally, we collected 15 detrital samples from four basins: Talara, Zorritos, Manabí, and Progreso. These rock units were previously mapped by the Geological and Cadastral Mining Information System of Peru and the Geological and Energy Research Institute of Ecuador. The distribution and nomenclature of the samples are detailed in Table 2.1 and Fig. 2.1.

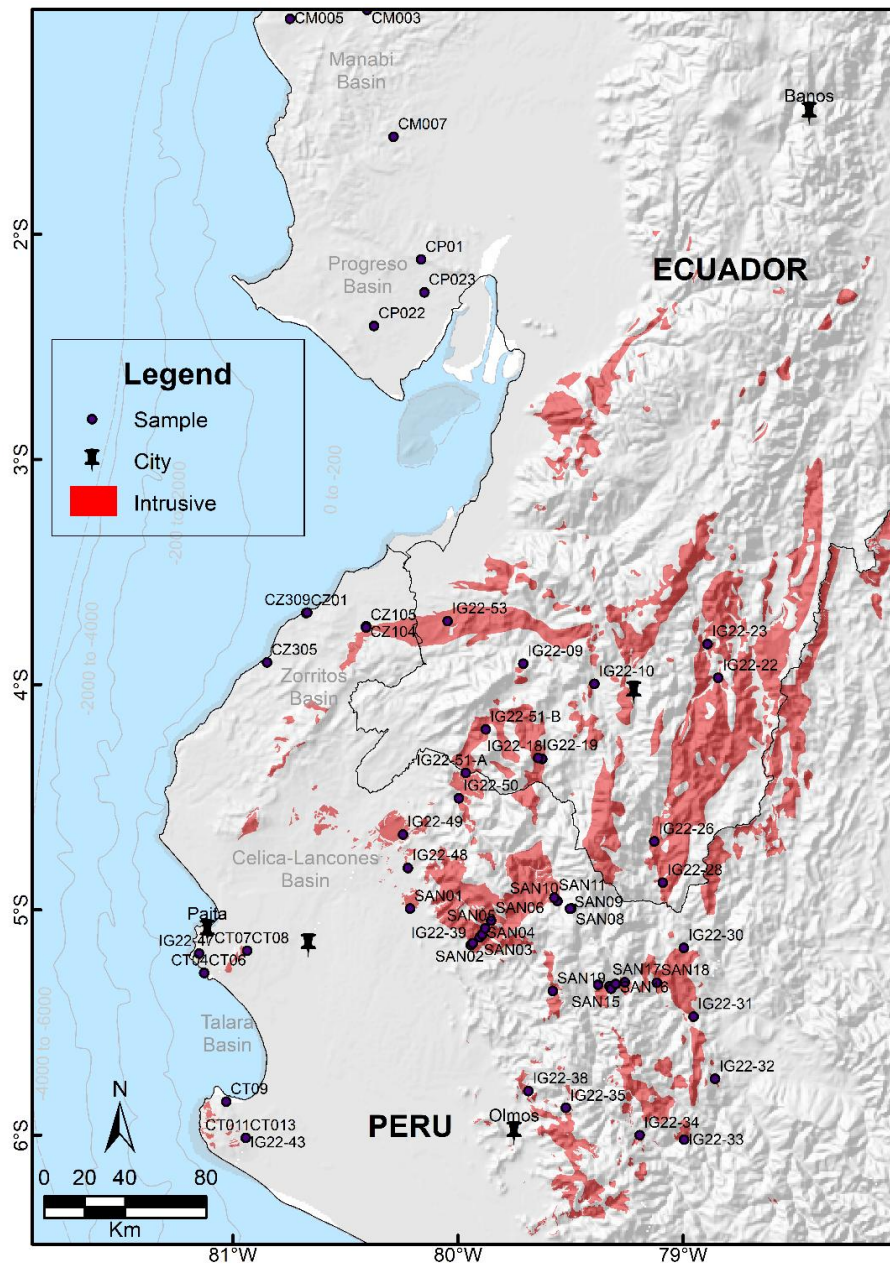


Figure 2.1 Map showing the location of the samples used in this project.

We prioritized selecting samples with minimal alteration at hand-sample size (Fig. 2.2). Each sample weighed between 2 and 4 kilograms, depending on its freshness. The collection took place during two field campaigns: October–November 2022 and January–February 2024, with the aim of sampling as many intrusive bodies as possible. However, some locations were inaccessible due to road conditions, weather, or community restrictions.

Table 2.1 Sample location

Name	Longitude	Latitude	Type	Name	Longitude	Latitude	Type
CM003	-80.4037	-1.0046	Detrital	IG22-32	-78.8593	-5.7508	Diorite
CM005	-80.7463	-1.0446	Detrital	IG22-33	-78.9964	-6.0205	Granodiorite
CM007	-80.2861	-1.5679	Detrital	IG22-34	-79.1937	-6.0013	Granodiorite
CM011	-80.2380	-0.9304	Detrital	IG22-35	-79.5213	-5.8796	Granodiorite
CM016	-80.1067	-0.8390	Detrital	IG22-38	-79.6880	-5.8051	Granodiorite
CP01	-80.1637	-2.1128	Detrital	IG22-39	-79.9434	-5.1569	Quartzmonzonite
CP022	-80.3730	-2.4081	Detrital	IG22-43	-80.9424	-6.0131	Metaquartzite
CP023	-80.1491	-2.2590	Detrital	IG22-47	-81.1494	-5.1949	Granodiorite
CT011	-80.9425	-6.0130	Metaquartzite	IG22-48	-80.2225	-4.8148	Granodiorite
CT013	-80.9425	-6.0130	Granite	IG22-49	-80.2444	-4.6660	Granodiorite
CT04	-81.1267	-5.2807	Detrital	IG22-50	-79.9967	-4.5051	Granodiorite
CT06	-81.1267	-5.2807	Detrital	IG22-51-A	-79.9659	-4.3925	Granodiorite
CT07	-80.9361	-5.1822	Detrital	IG22-51-B	-79.8767	-4.1992	Diorite
CT08	-80.9361	-5.1822	Detrital	IG22-53	-80.0464	-3.7183	Granite
CT09	-81.0292	-5.8520	Metaquartzite	SAN01	-80.2131	-4.9948	Granitoid
CZ01	-80.6740	-3.6829	Detrital	SAN02	-79.9345	-5.1487	Granite
CZ104	-80.4084	-3.7418	Granite	SAN03	-79.9043	-5.1225	Granite
CZ105	-80.4078	-3.7469	Granite	SAN04	-79.8928	-5.1099	Granodiorite
CZ305	-80.8480	-3.9020	Detrital	SAN05	-79.8795	-5.0830	Granodiorite
CZ309	-80.6697	-3.6806	Detrital	SAN06	-79.8528	-5.0485	Granitoid
IG22-09	-79.7103	-3.9074	Metaquartzite	SAN08	-79.4967	-4.9940	Granodiorite
IG22-10	-79.3942	-3.9973	Metaquartzite	SAN09	-79.5030	-4.9956	Granodiorite
IG22-18	-79.6262	-4.3300	Gabbro	SAN10	-79.5585	-4.9614	Granodiorite
IG22-19	-79.6444	-4.3260	Granodiorite	SAN11	-79.5725	-4.9467	Granitoid
IG22-22	-78.8448	-3.9704	Granodiorite	SAN13	-79.3777	-5.3329	Granite
IG22-23	-78.8921	-3.8204	Granodiorite	SAN14	-79.3271	-5.3395	Granite
IG22-26	-79.1287	-4.6966	Granodiorite	SAN15	-79.3205	-5.3510	Granite
IG22-28	-79.0907	-4.8796	Quartzmonzonite	SAN16	-79.2997	-5.3300	Granitoid
IG22-30	-78.9973	-5.1693	Granite	SAN17	-79.2593	-5.3223	Granitoid
IG22-31	-78.9545	-5.4746	Granodiorite	SAN18	-79.1176	-5.3246	Granitoid
				SAN19	-79.5791	-5.3608	Granodiorite

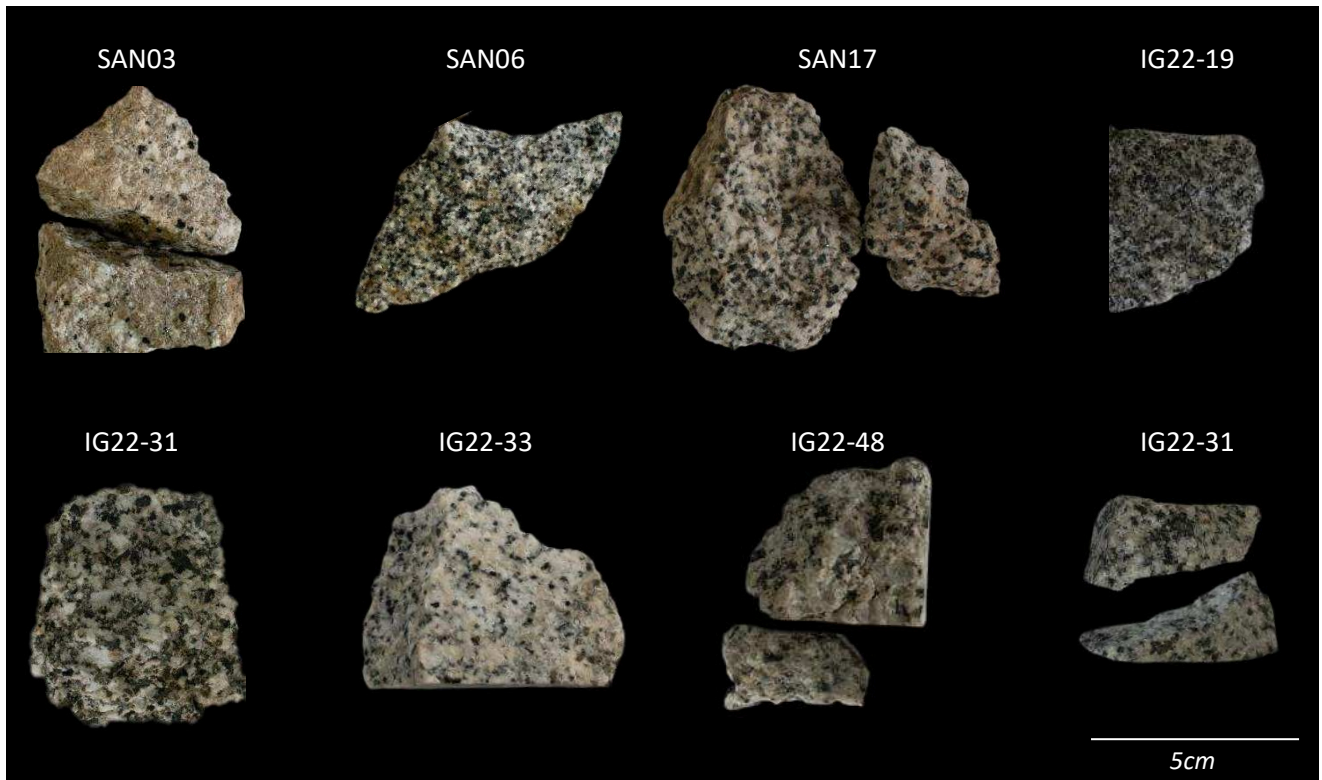


Figure 2.2 Hand sample images of representative studied rocks.

2.2 Sample preparation

Sample preparation took place at the Pedagogical and Technological University of Colombia UPTC, Sogamoso Sectional, in the thermochronology laboratory, following standard techniques (Kohn et al., 2019), which summarized method is illustrated in Fig. 2.3. Initially, samples underwent brushing, washing, and natural drying for a day to eliminate any external materials such as mud or biological residues on their surfaces. Subsequently, samples were crushed with a cast iron jaw crusher to reduce them to fragments smaller than 5 centimeters in size. Grinding was then carried out using a tungsten alloy disk mill while ensuring that crystals did not become too small for practical applications, aiming for sizes <1 mm. Subsequently, samples underwent sieving using an electric automatic sieve shaker to separate materials with size <500 μm . Hydro-gravimetric separation was employed to retain the heavy fraction of the samples utilizing an automated gold spiral wheel machine, guaranteeing that excessive sample material was not lost during this step. Magnetic separation was performed using hand magnets and a Frantz magnetic separator, with current increments ranging from 0.2 microamperes to 1 microampere, to isolate the magnetic components of the sample. Subsequently, gravimetric separation was carried out using standard heavy liquids (bromoform: 2.96 g/cm^3 and diiodomethane: 3.33 g/cm^3) to separate zircons and apatites from the non-magnetic fraction.

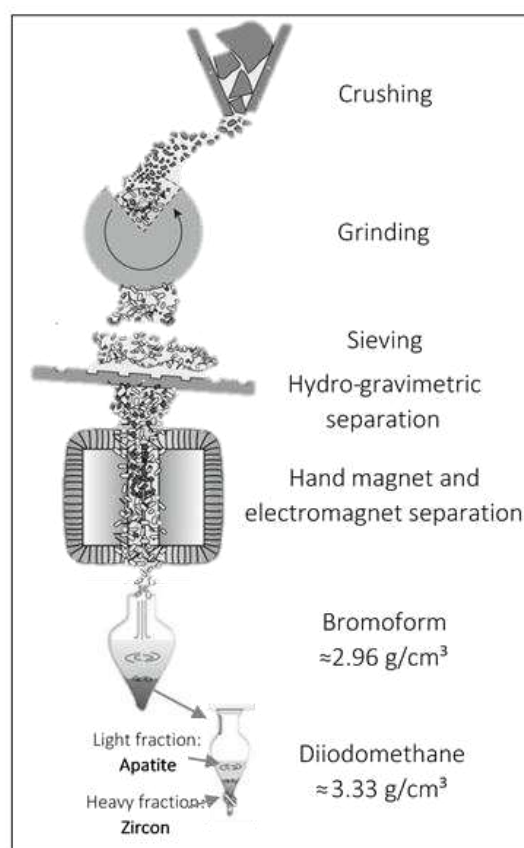


Figure 2.3 Generalized sequence of steps for separation of apatite and zircon minerals, modified from Kohn et al. (2019). Additionally, portions of the crushed sample were reserved for thin-section preparation, while powdered fractions were retained for whole-rock ICP-AES analysis.

2.3 U-Pb methodology

Main dating techniques used in geology and petrology rely on radiometric methods, which involve measuring the ratios of parent-daughter isotopes. In principle, the concentrations of these isotopes change and accumulate over time, once radioactive equilibrium is established (Geyh & Schleicher, 1990). Among other methods, U-Pb dating provides access to two distinct geochronometers ($^{206}\text{Pb}/^{238}\text{U}$ and $^{207}\text{Pb}/^{235}\text{U}$) that rely on different isotopes of the same parent-daughter pair (U and Pb). This inherent duplication serves as a robust internal verification, rendering this method one of the most reliable dating techniques (Schoene, 2014). In this particular case, if a mineral begins to accumulate radiogenic Pb in a closed system, and $^{207}\text{Pb}/^{235}\text{U}$ and $^{206}\text{Pb}/^{238}\text{U}$ evolve in such way that both ratios follow the concordia curve, the derived age is considered reliable and can be used, for example, to determine the crystallization age or time of a high-temperature event of a mineral.

Certain minerals, such as zircon, have the ability to absorb significant quantities of uranium (U) and very little lead (Pb) during their formation, making them ideal for geochronological studies.

Few common Pb in zircon allows the simplification of general U-Pb isochron equations to equations 2.1 and 2.2 (Hoskin & Schaltegger, 2003).

$$\frac{{}^{206}\text{Pb}}{{}^{238}\text{U}} = (e^{\lambda_{238}t} - 1) \quad \text{equation 2.1}$$

$$\frac{{}^{207}\text{Pb}}{{}^{235}\text{U}} = (e^{\lambda_{235}t} - 1) \quad \text{equation 2.2}$$

Where t is the time since the system closed, and λ_{238} and λ_{235} , are the decay constants of ${}^{238}\text{U}$, ${}^{235}\text{U}$, respectively.

Sixty-one zircon mounts, were prepared at Lille University. Zircon picking was performed using an Olympus SZX16 binocular microscope, and mounted on araldite Teflon sheets. The samples were polished on a lap wheel using successive diamond suspensions of 9 μm , 6 μm , and 3 μm , until the crystal surfaces were revealed.

Cathodoluminescence (CL) imaging (Fig. 2.4) was performed in order to recognize different micro-area compositional heterogeneities within a crystal before U-Pb dating (Kreidler, 1985). CL imagen acquisition was performed using a cold cathodoluminescence model 8200 MkII equipment mounted in an Olympus BX41 microscope at Lille university.

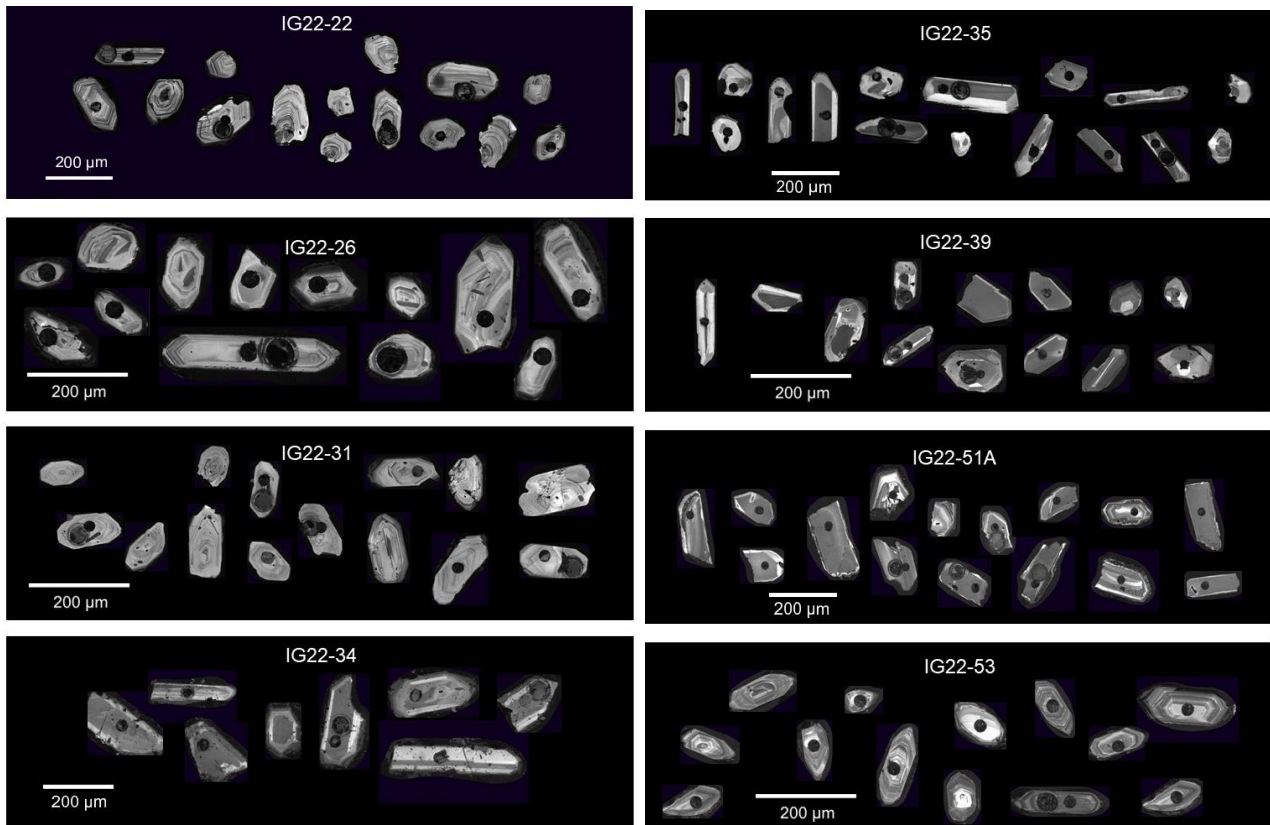


Figure 2.4 Cathodoluminescence images of selected samples

Zircon U-Pb dating was conducted using a LA-ICP-MS at the GeOHeLiS analytical platform (OSUR, Univ. Rennes) using an ESI NWR193UC Excimer laser coupled to a quadrupole Agilent 7700x ICP-MS equipped with a dual pumping system to enhance sensitivity. Ablated material was carried into Helium mixed with nitrogen and argon, followed by injection into the plasma source. Mass calibration was performed before each session using the NIST SRM 612 reference glass to inspect the ^{238}U signal. The analysis involved acquiring signals of $^{204}(\text{Pb}+\text{Hg})$, ^{206}Pb , ^{207}Pb , ^{208}Pb and ^{238}U masses, during single analyses, which consisted of background detection for 20 s, followed by 60 s of sample measurement, and then 15 s delay for sample washout. Ablation spot diameters ranged from 25– 50 μm depending on grain size. Repeated measurements of the GJ-1 (Jackson et al., 2004) zircon standard were used for U-Pb and Th-Pb fractionation correction as well as mass bias reduction. Plešovice (Sláma et al., 2008) and 91500 (Goolaerts et al., 2004) zircon standards were utilized as measurements to ensure precision and accuracy of the analysis. Data reduction process utilized the GLITTER® software package developed by Macquarie Research Ltd (van Achterberg et al., 2001). Additionally, data were analyzed using the ISOPLOT package program (Ludwig, 2003); $^{207}\text{Pb}/^{206}\text{Pb}$ dates were used for zircons grains older than 1Ga, whereas $^{206}\text{Pb}/^{238}\text{U}$ dates were used for grains younger than 1Ga. Errors are reported at 2 sigmas, and range from 90% to 110% concordance. Summarized U-Pb ages are given in table 2.2.

Table 2.2 U-Pb ages

Sample	Latitude	Longitude	Age U-Pb	$\pm 2\sigma$	Sample	Latitude	Longitude	Age U-Pb	$\pm 2\sigma$
CM003	-80.404	-1.005	604-619Ma		IG22-32	-78.859	-5.751	157.7	4.2
CM005	-80.746	-1.045	229-572Ma		IG22-33	-78.996	-6.020	35.9	0.2
CM007	-80.286	-1.568	220-1033Ma		IG22-34	-79.194	-6.001	187.8	3.2
CM011	-80.238	-0.930	462-682 Ma		IG22-35	-79.521	-5.880	32.7	0.3
CM016	-80.107	-0.839	234-1126 Ma		IG22-38	-79.688	-5.805	29.6	0.6
CP01	-80.164	-2.113	204-1654 Ma		IG22-39	-79.943	-5.157	40.7	0.3
CP022	-80.373	-2.408	214-3356 Ma		IG22-47	-81.149	-5.195	93.3	1.4
CP023	-80.149	-2.259	213-1078 Ma		IG22-48	-80.223	-4.815	97.2	1.4
CT011	-80.943	-6.013	340-3165 Ma		IG22-49	-80.244	-4.666	87.8	1.7
CT04	-81.127	-5.281	289- 2715 Ma		IG22-50	-79.997	-4.505	97.6	1.4
CT06	-81.127	-5.281	306-2644 Ma		IG22-51-A	-79.966	-4.392	99.1	2.3
CT07	-80.936	-5.182	322-2479 Ma		IG22-51-B	-79.877	-4.199	100.8	1.8
CT08	-80.936	-5.182	224-766 Ma		IG22-53	-80.046	-3.718	235.5	3.7
CT09	-81.029	-5.852	308-2716 Ma		SAN01	-80.213	-4.995	90.0	2.8
CZ01	-80.674	-3.683	204-1654 Ma		SAN02	-79.935	-5.149	45.2	0.3
CZ305	-80.848	-3.902	215-2390 Ma		SAN03	-79.904	-5.122	44.1	0.3
CZ309	-80.670	-3.681	222-1111 Ma		SAN04	-79.893	-5.110	46.7	0.4
IG22-09	-79.710	-3.907	231-801 Ma		SAN05	-79.880	-5.083	42.0	0.3
IG22-10	-79.394	-3.997	619-1045 Ma		SAN06	-79.853	-5.049	43.6	0.7
IG22-43	-80.942	-6.013	265-1584 Ma		SAN08	-79.497	-4.994	20.2	0.3
CT13	-80.943	-6.013	222.8	1.2	SAN09	-79.503	-4.996	20.2	0.4

Table 2.2: Continuete.

Sample	Latitude	Longitude	Age U-Pb	$\pm 2\sigma$
CZ104	-80.408	-3.742	222	1.2
CZ105	-80.408	-3.747	229.2	1.4
IG22-18	-79.626	-4.330	94.9	1.4
IG22-19	-79.644	-4.326	91.4	1.5
IG22-22	-78.845	-3.970	186.1	6.5
IG22-23	-78.892	-3.820	191.4	5.3
IG22-26	-79.129	-4.697	179.2	1.2
IG22-28	-79.091	-4.880	137.5	5.8
IG22-30	-78.997	-5.169	126.1	3.6
IG22-31	-78.954	-5.475	133.7	0.9

Sample	Latitude	Longitude	Age U-Pb	$\pm 2\sigma$
SAN10	-79.558	-4.961	29.1	0.4
SAN11	-79.573	-4.947	58.9	0.6
SAN13	-79.378	-5.333	34.3	0.4
SAN14	-79.327	-5.339	33.7	0.4
SAN15	-79.320	-5.351	33.7	0.4
SAN16	-79.300	-5.330	33.6	0.4
SAN17	-79.259	-5.322	40.2	0.4
SAN18	-79.118	-5.325	139.8	1.8
SAN19	-79.579	-5.361	19.9	0.4

2.4 Petrographic analyses

Twenty-four thin sections were prepared at the University of Padova. The rocks were cut with a rock saw to obtain small rectangular parallelepipeds with flat surfaces. These surfaces were then polished by hand using 500 and 800 grit abrasive powders to ensure smoothness and eliminate surface roughness. The polished samples were mounted on glass slides using epoxy resin. After mounting, the samples were ground and polished both manually and using an automated rotating wheel with progressively finer abrasives (600 to 1000 grits) to achieve a final thickness of approximately 30 μm . Thickness was verified by observing the interference colors of quartz under polarized light using a petrographic microscope.

Each thin section was analyzed using a petrographic microscope under both plane-polarized light (PPL) and cross-polarized light (XPL) at magnifications of 5 $\times$, 10 $\times$ and 20 $\times$ to identify mineral assemblages, textures and grain shapes. Modal mineralogy was quantified using a point-counting method along an S-shaped traverse, recording 200 points per thin section by identifying the mineral at the center of the eyepiece reticle. Photographs of the thin sections were also taken to document textures and mineral assemblages. The mineralogical compositions are summarized in Table 2.3 and representative images of the samples are shown in Fig. 2.5.

Table 2.3 Petrography analysis results of the samples from this study

Sample	Rock-forming minerals						Secondary minerals	Accessory minerals					Texture	General grain shape	
	Qz	Alkali Feldspar			Plagioclase			Msv	Bt	Apt	Hrb	Cpx			Opq
		Mc	Or	Sn	Ab	An									
IG22-19	9	0	10	0	0	46	17	0	0	18	0	0	Phaneritic	Euhedral to subhedral	
IG22-22	5	17	8	0	21	30	8	0	0	6	0	5	Phaneritic	Euhedral to subhedral	
IG22-23	12	12	8	0	15	31	16	0	6	0	0	0	Phaneritic	Euhedral to subhedral	
IG22-28	18	28	0	0	28	14	8	0	4	0	0	0	Phaneritic	Euhedral to subhedral	
IG22-30	30	16	12	0	10	26	6	0	0	0	0	0	Phaneritic	Euhedral to subhedral	
IG22-31	12	0	15	0	7	35	9	0	6	9	0	7	Phaneritic	Euhedral to subhedral	
IG22-32	6	19	0	0	42	0	18	0	6	9	0	0	Phaneritic	Euhedral to subhedral	
IG22-33	22	18	9	0	16	26	9	0	0	0	0	0	Phaneritic	Euhedral to subhedral	
IG22-38	18	13	5	0	16	26	12	10	0	0	0	0	Phaneritic	Euhedral to subhedral	
IG22-39	15	7	10	0	26	32	0	0	4	6	0	0	Phaneritic	Euhedral to subhedral	
IG22-48	17	19	10	0	35	0	19	0	0	0	0	0	Phaneritic	Euhedral to subhedral	
IG22-49	17	17	0	0	29	0	25	0	0	10	0	2	Phaneritic	Euhedral to subhedral	
IG22-50	12	0	7	0	23	38	8	0	0	9	0	3	Phaneritic	Euhedral to subhedral	
IG22-51-A	8	0	0	0	20	38	16	0	7	11	0	0	Phaneritic	Euhedral to subhedral	
SAN01	24	0	18	0	0	33	21	0	0	0	0	4	Phaneritic	Euhedral to subhedral	
SAN03	17	13	9	0	12	18	18	5	3	5	0	0	Phaneritic	Euhedral to subhedral	
SAN08	30	22	0	0	0	37	0	7	4	0	0	0	Phaneritic	Euhedral to subhedral	
SAN09	19	13	0	0	22	31	0	10	5	0	0	0	Phaneritic	Euhedral to subhedral	
SAN10	16	10	0	0	30	39	0	0	5	0	0	0	Phaneritic	Euhedral to subhedral	
SAN06	17	0	0	8	0	44	14	7		10	6	0	Phaneritic	Euhedral to subhedral	
SAN13	18	10	11	0	21	29	11	0	0	0	0	0	Phaneritic	Euhedral to subhedral	
SAN14	21	19	16	7	11	15	11	0	0	0	0	0	Phaneritic	Euhedral to subhedral	
SAN19	24	24	0	0	40	0	0	12	0	0	0	0	Phaneritic	Euhedral to subhedral	
SAN05	16	13	0	0	0	43	0	8	4	9	7	0	Phaneritic	Euhedral to subhedral	

Abbreviations: Qz: Quartz. Mc: microcline. Or: orthoclase. Sn: sanidine. Ab: albite. An: anorthite. Msv: muscovite. Bt: biotite. Apt: apatite. Hrb: hornblende. Cpx: clinopyroxene. Opq. Opaque.

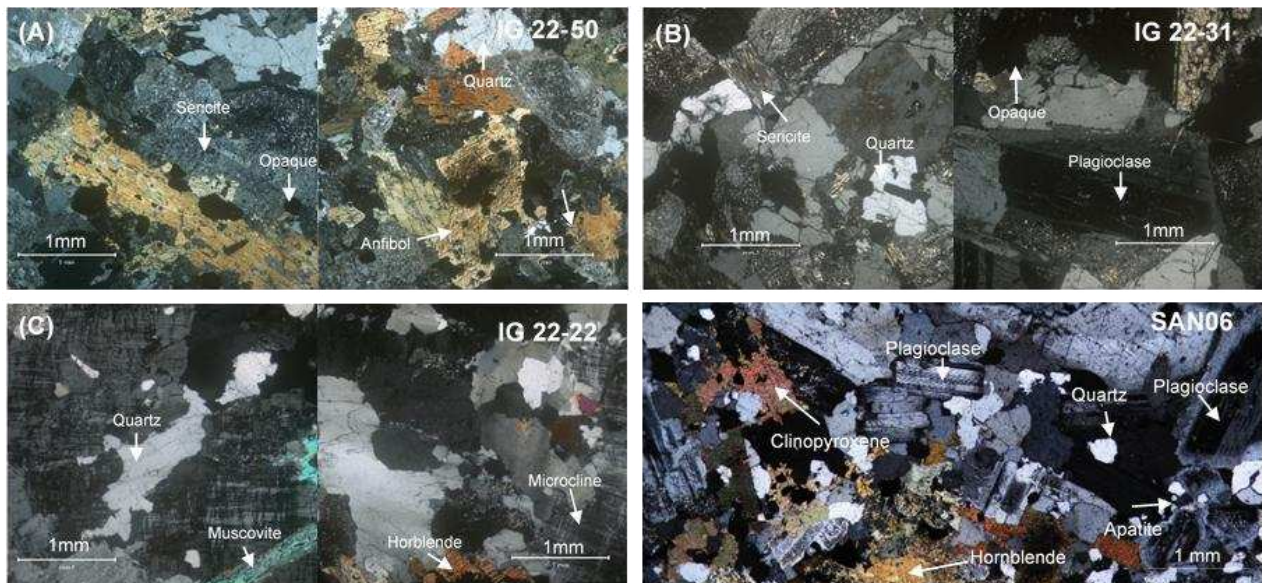


Figure 2.5 Photomicrographs of some representative samples displaying main mineral assemblages

2.5 Whole rock chemistry

The elemental compositions of twenty-seven powder samples, were analyzed at the Institut Universitaire Européen de la Mer, Université de Bretagne Occidentale (Brest, France) using a Jobin Yvon Ultima 2 inductively coupled plasma atomic emission spectrometer (ICP-AES). Sample preparation followed the analytical procedure described by Cotten et al. (1995), including digestion with HF and HNO₃, neutralization with boric acid and subsequent dilution in nitric acid. Calibration was performed using international standards (AC-E, BEN, JB-2, PM-S and WS-E), together with internal standards (MORB-E, CB2, CB15 and CB18), ensuring external reproducibility of between 5 and 10%, depending on the element and concentration. The relative standard deviations were $\pm 1\%$ for SiO₂ and $\pm 2\%$ for other major elements, except for P₂O₅ and MnO, which had an absolute precision of $\pm 0.01\%$. Results are shown in Table 2.3.

Table 2.4 Whole-Rock composition of major and trace elements analyzed by ICP-AES

	SAN 01	SAN 02	SAN 03	SAN 05	SAN 08	SAN 09	SAN 10	SAN 13	SAN 15	SAN 19	IG22-33	IG22-38	IG22-39
SiO ₂	69.02	75.62	78.11	64.81	67.30	66.45	64.68	76.24	73.54	64.10	69.70	63.82	53.52
TiO ₂	0.44	0.14	0.13	0.52	0.55	0.47	0.68	0.14	0.20	0.61	0.26	0.63	1.09
Al ₂ O ₃	14.40	12.33	11.00	14.97	15.19	15.57	15.22	12.89	13.77	15.37	14.33	16.52	18.95
Fe ₂ O ₃	4.09	1.58	1.51	4.97	4.93	4.42	5.06	1.76	2.19	6.05	3.50	5.15	9.43
MnO	0.08	0.03	0.03	0.07	0.08	0.08	0.05	0.07	0.05	0.11	0.11	0.09	0.19
MgO	1.10	0.16	0.20	1.99	1.99	1.81	1.35	0.21	0.47	2.28	0.87	2.09	3.69
CaO	2.67	1.11	0.87	4.76	4.30	4.64	2.92	0.39	0.38	3.53	3.64	5.01	7.76
Na ₂ O	3.67	3.85	3.24	3.34	3.10	3.17	3.89	3.68	3.42	2.73	3.87	3.93	3.75
K ₂ O	2.03	3.36	4.09	2.05	3.00	2.64	2.59	4.21	5.17	3.34	2.36	1.38	0.78
P ₂ O ₅	0.12	0.02	0.03	0.09	0.10	0.09	0.18	0.04	0.03	0.10	0.13	0.17	0.18
LOI	1.68	0.45	0.47	0.65	0.93	0.51	2.61	1.31	2.43	1.62	1.42	2.52	1.15
Rb	39.47	112.70	105.17	54.79	97.30	84.16	60.99	106.50	147.20	133.55	55.58	27.71	25.72
Sr	1273.53	51.63	34.67	181.69	233.72	256.57	248.07	65.45	73.88	251.39	92.51	445.58	292.09
Ba	206.54	718.61	536.51	443.64	518.27	458.07	641.69	369.54	376.24	614.44	760.75	531.29	224.94
Sc	6.43	3.80	3.20	13.51	12.29	11.45	11.83	5.30	5.47	14.36	3.59	9.84	30.64
Co	2.76	<LD	<LD	9.13	8.26	7.58	5.91	0.01	0.36	10.08	2.88	9.47	15.80
Ni	10.69	2.18	1.73	5.04	9.49	9.01	4.06	3.39	1.70	10.06	3.85	12.00	6.86
Y	2.49	10.97	16.69	22.62	22.56	19.66	26.37	19.23	22.62	23.56	16.23	9.65	23.06
Nb	4.95	5.10	2.77	4.68	4.63	3.70	6.21	9.09	8.25	5.19	6.12	5.95	4.72
La	65.80	11.62	19.79	10.44	15.41	14.61	23.17	16.07	13.84	12.18	8.47	18.09	9.24
Ce	9.81	25.88	38.76	24.16	36.29	33.12	48.10	39.79	32.36	32.35	15.60	39.75	21.88
Nd	23.73	18.96	26.82	25.87	28.42	27.19	37.87	28.62	25.23	28.50	8.64	18.00	13.07
Sm	24.32	1.69	2.82	3.55	4.09	3.45	5.69	3.88	3.60	4.03	2.41	3.62	3.00
Eu	2.93	0.42	0.23	0.83	0.69	0.79	1.15	0.36	0.34	0.76	0.88	1.14	1.15
Gd	0.88	2.04	2.59	3.36	3.90	3.58	5.51	4.19	3.68	4.15	1.83	2.39	3.37
Dy	3.48	1.66	2.32	3.58	3.56	3.30	4.58	3.28	3.39	3.84	2.12	1.90	3.72
Er	3.67	0.57	1.14	2.36	2.15	1.87	2.47	1.58	1.70	1.97	1.79	0.44	1.67
Yb	2.29	1.09	1.60	2.16	1.93	1.77	2.13	1.88	2.28	1.98	1.56	0.66	2.16
Th	2.64	8.29	10.03	6.78	12.80	11.62	8.43	12.90	16.01	11.49	6.33	3.29	1.34

Table 2.4 Continued

	IG22-18	IG22-19	IG22-22	IG22-23	IG22-26	IG22-28	IG22-30	IG22-31	IG22-47	IG22-48	IG22-49	IG22-50	IG22-51B	IG22-53
SiO ₂	48.2	62.9	67.6	64.2	64.1	63.0	71.0	64.0	73.7	70.4	73.7	65.5	58.9	71.8
TiO ₂	0.6	0.3	0.4	0.5	0.6	0.6	0.3	0.5	0.2	0.4	0.3	0.6	0.6	0.4
Al ₂ O ₃	17.8	14.8	16.2	16.4	16.7	16.9	14.7	15.2	13.9	14.2	13.6	14.6	19.2	13.1
Fe ₂ O ₃	7.9	4.0	4.5	5.8	5.9	6.7	2.4	5.8	1.3	3.3	2.6	4.7	4.4	2.9
MnO	0.2	0.1	0.1	0.1	0.1	0.2	0.0	0.1	0.0	0.1	0.1	0.1	0.1	0.0
MgO	7.6	3.8	1.6	2.0	1.7	1.9	0.5	2.1	0.2	0.9	0.6	1.7	1.7	0.7
CaO	13.8	7.8	3.4	5.0	5.4	2.5	2.4	5.1	0.6	3.0	2.3	4.5	7.1	1.1
Na ₂ O	1.7	2.5	3.7	3.5	3.6	4.9	3.2	2.9	2.8	3.9	4.0	4.5	5.5	2.6
K ₂ O	0.3	1.8	2.5	2.0	1.6	2.6	4.5	3.2	5.1	1.9	2.8	0.7	0.7	4.2
P ₂ O ₅	0.1	0.1	0.1	0.1	0.2	0.1	0.1	0.2	0.3	0.1	0.1	0.1	0.1	0.1
LOI	1.9	0.8	1.4	1.6	1.9	1.8	1.2	1.9	1.1	1.5	1.3	1.7	1.7	1.9
Rb	8.3	37.2	59.2	77.3	39.7	76.8	137.9	57.1	261.5	41.5	70.6	19.1	12.0	118.0
Ba	118.8	1459.7	688.7	507.5	543.5	1702.0	1043.3	1203.7	338.1	1340.2	1131.1	499.0	503.3	463.8
Sr	310.4	274.0	267.2	285.9	329.6	258.6	251.4	546.4	52.2	204.3	178.9	221.7	254.7	83.6
Th	1.5	7.3	8.5	12.4	4.8	10.0	44.8	12.5	6.4	6.3	10.2	5.1	4.3	7.7
Zr	38.0	70.9	49.3	58.6	41.1	81.7	56.7	32.1	79.6	44.6	102.9	62.9	36.0	39.1
Y	12.4	12.8	12.9	27.6	17.3	25.7	16.0	14.8	11.6	24.4	25.0	32.0	42.9	12.9
Nb	0.5	0.8	2.7	4.4	4.1	5.7	4.3	4.0	13.3	3.1	3.6	1.2	0.8	7.2
Sc	53.4	25.4	8.5	15.5	9.0	16.5	4.3	12.9	2.7	9.1	7.4	19.6	21.7	6.2
Cr	366.7	144.4	8.0	7.5	8.5	8.7	13.2	9.3	5.4	8.0	7.5	11.0	15.9	22.2
Ni	45.9	38.2	4.3	6.1	4.0	9.6	8.7	7.4	3.1	6.2	5.0	7.0	6.3	11.1
Co	27.8	13.7	7.0	10.6	8.7	10.2	1.7	11.9	<LD	3.3	1.3	6.6	15.1	4.1
V	223.7	91.1	82.6	122.1	72.4	123.5	32.3	138.9	10.4	44.0	31.1	101.0	102.7	33.9
La	7.3	10.3	12.0	18.1	18.5	23.3	14.3	18.8	15.4	9.7	20.1	11.1	11.6	19.6
Ce	16.9	21.3	20.0	41.7	37.7	50.8	28.5	40.7	31.3	20.0	42.8	28.2	27.9	43.6
Nd	7.9	10.6	9.7	20.1	19.8	25.3	13.3	20.1	14.5	10.6	18.2	15.7	17.0	19.9
Sm	1.9	1.9	2.0	4.6	4.5	5.5	2.7	4.4	3.7	2.9	3.5	4.5	4.6	4.2
Eu	0.5	0.6	0.7	1.0	1.0	0.9	0.8	1.1	0.5	0.8	0.7	1.0	1.3	0.7
Gd	2.2	1.5	1.9	4.5	3.8	4.7	2.9	3.3	3.8	3.1	3.7	4.9	5.6	4.3
Dy	1.9	1.9	1.8	4.2	2.9	4.2	2.3	2.2	2.3	3.4	3.9	4.8	6.5	2.7
Er	1.0	1.2	1.3	2.7	1.6	2.5	1.5	1.1	0.3	2.0	2.5	2.8	4.1	0.8
Yb	1.1	1.4	1.3	2.7	1.4	2.4	1.5	1.3	0.4	2.5	2.7	3.0	4.4	0.7

2.6 Zircon chemistry

2.6.1 Lu-Hf system

Hafnium is more incompatible than lutecium during the melting of spinel and garnet peridotite (Patchett & Tatsumoto, 1980), resulting in a long-term enrichment of Hf compared to Lu in the continental crust, leading to relatively un-radiogenic crust and radiogenic mantle reservoirs (Patchett et al., 1982). Therefore, the ^{176}Lu – ^{176}Hf system can be used to constrain the development of the mantle-crust system throughout Earth's geological history (Vervoort & Patchett, 1996).

Hafnium isotopic data is typically expressed as the $^{176}\text{Hf}/^{177}\text{Hf}$ ratio, which can vary depending on the sample's age and origin. To account for initial heterogeneity in hafnium isotopes and focus on variations driven by geological processes, researchers use the epsilon Hf ($\epsilon\text{Hf}_{(t)}$) notation (Equation 2.3) for normalization (Faure & Mensing, 2005). This process involves measuring the $^{176}\text{Lu}/^{177}\text{Hf}$ ratio of a sample and comparing it to a standard reference material, CHUR (Chondritic Uniform Reservoir), which represents the isotopic composition of the primitive mantle (Bouvier et al., 2008).

$$\epsilon\text{Hf}_0 = \left[\frac{\left(\frac{^{176}\text{Hf}}{^{177}\text{Hf}} \right)_{\text{sample}(0)}}{\left(\frac{^{176}\text{Hf}}{^{177}\text{Hf}} \right)_{\text{CHUR}(0)}} \right] * 10000 \quad \text{equation 2.3}$$

Hafnium is a lithophile element with chemical properties similar to zirconium. Zircon, a zirconium silicate mineral, preserves the initial $^{176}\text{Hf}/^{177}\text{Hf}$ ratio of the magma at the time of its formation. This characteristic makes zircon an invaluable mineral for studying the Lu-Hf isotopic system, providing a key marker for understanding crustal evolution (Hawkesworth & Kemp, 2006).

Lu-Hf isotope analyses were performed at the Géosciences Montpellier Laboratory, University of Montpellier, France (AETE-ISO regional facility of OSU OREME), using a ThermoFinnigan Neptune+ MC-ICP-MS (multicollector inductively coupled plasma-mass spectrometer) coupled to a Photon-Machine Analyte G2 excimer laser (193 nm wavelength). Analyses were performed over two data collection periods. To ensure data quality and reproducibility, three zircon standards - 91500, GJ1 and Plešovice (with reference values from: Blichert-Toft, 2008; Morel et al., 2008; Sláma et al., 2008), were analyzed as unknowns in both sessions.

In the first session, ablation was performed with a spot size of 50 μm , a laser frequency of 5 Hz and an energy density of 5 J/cm². The values obtained were in agreement with the reference standards, with 91500 yielding a $^{176}\text{Hf}/^{177}\text{Hf}$ ratio of 0.282305 ± 0.000020 ($n = 46$), GJ1 yielding 0.282015 ± 0.000027 ($n = 45$), and Plešovice yielding 0.282483 ± 0.000100 ($n = 45$). During the second session, the laser was operated at a repetition rate of 6 Hz, with the same spot size of 50 μm , but a slightly lower fluence of 4.5 J/cm². The results were again in agreement with the reference standards, with 91500 giving a $^{176}\text{Hf}/^{177}\text{Hf}$ ratio of 0.282305 ± 0.000020 , GJ1 giving 0.282005 ± 0.000036 and Plešovice giving 0.282486 ± 0.000010 . This consistency across both sessions demonstrates the robustness and reproducibility of the analytical procedure. A total of 387 new Lu-Hf measurements were collected, with the data presented in Table 2.5 and illustrated in Fig. 2.6.

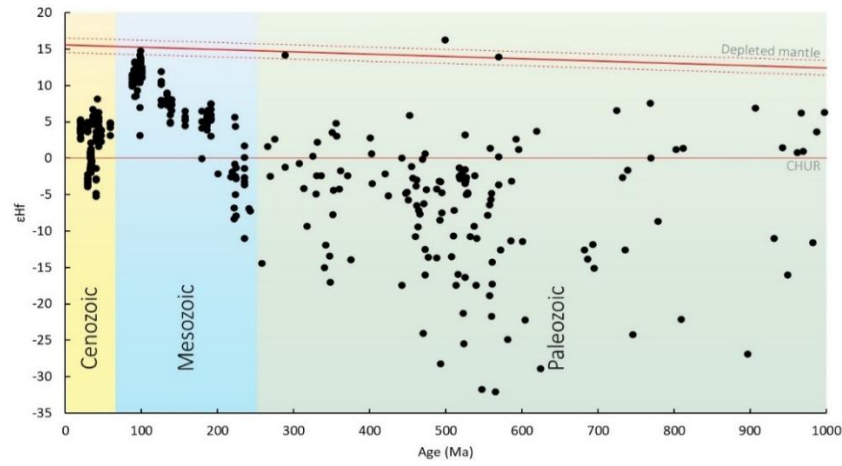


Figure 2.6 ϵ_{Hf} vs. Age isotopic data on zircon

Table 2.5 zircon ϵ_{Hf} data

Sample	Analysis	Age (Ma)	ϵ_{Hf}	Sample	Analysis	Age (Ma)	ϵ_{Hf}	Sample	Analysis	Age (Ma)	ϵ_{Hf}
CT010	ge_1	350.6	3.5	CT004	ge_31	494.9	-4.7	CT006	ge_62	456.7	-2.7
CT010	ge_2	442.6	0.1	CT004	ge_32	487.6	-4.2	CT006	ge_63	449.2	-4.7
CT010	ge_3	455.3	-1.2	CT004	ge_33	942.3	1.5	CT009	ge_64	223.8	4.4
CT010	ge_4	356	4.8	CT004	ge_34	1081.7	2.5	CT009	ge_65	307.0	-0.7
CT010	ge_5	949.3	-16.0	CT004	ge_35	519.1	-2.8	CT009	ge_66	779.0	-8.7
CT010	ge_6	525.7	-3.4	CT004	ge_36	465.9	-7.6	CT009	ge_67	569.4	0.2
CT010	ge_7	559.7	-4.8	CT004	ge_37	906.9	6.9	CT009	ge_68	463.2	-9.4
CT010	ge_8	424.3	-5.2	CT004	ge_38	289.1	14.2	CT009	ge_69	491.1	-3.2
CT010	ge_9	560.7	-14.3	CT004	ge_39	738.9	-1.7	CT009	ge_70	1026.6	-1.5
CT010	ge_10	402.3	0.6	CT004	ge_40	802.4	1.2	CT009	ge_71	528.6	-4.7
CT010	ge_11	1069	8.0	CT004	ge_41	997.4	6.3	CT009	ge_72	469.5	-0.1
CT010	ge_12	331.2	2.2	CT004	ge_42	400.9	2.8	CT009	ge_73	1050.5	4.5
CT010	ge_13	526.6	-2.6	CT004	ge_43	526	-3.1	CT009	ge_74	472.7	0.6
CT010	ge_14	525.2	-1.5	CT006	ge_44	474.1	-4.3	CT009	ge_75	471.2	-6.2
CT010	ge_15	360	-4.2	CT006	ge_45	464.2	-7.1	CT009	ge_76	569.3	-3.6
CT010	ge_16	329.8	-4.9	CT006	ge_46	461.0	-3.8	CT009	ge_77	336.3	-2.4
CT010	ge_17	510.9	-7.2	CT006	ge_47	317.2	-9.3	CT009	ge_78	495.0	-7.5
CT010	ge_18	352	-4.4	CT006	ge_48	447.1	-4.8	CZ018	ge_79	371.1	-2.4
CT010	ge_19	528.7	-4.9	CT006	ge_49	461.5	-6.4	CZ018	ge_80	222.3	5.7
CT010	ge_20	559.5	-5.7	CT006	ge_50	517.6	-2.5	CZ018	ge_81	288.4	-1.2
CT010	ge_21	441.9	-17.4	CT006	ge_52	518.3	-2.8	CZ018	ge_82	361.9	-1.7
CT010	ge_22	527.3	-5.0	CT006	ge_53	477.2	-13.6	CZ018	ge_83	241.5	-6.9
CT010	ge_23	557.3	-6.4	CT006	ge_54	769.3	0.1	CZ018	ge_84	452.1	5.9
CT010	ge_24	1078.7	9.1	CT006	ge_55	459.8	-10.8	CZ018	ge_85	420.5	-2.2
CT010	ge_25	560.2	-21.7	CT006	ge_56	330.4	-2.4	CZ018	ge_98	571.9	-12.6
CT010	ge_26	520.7	-2.0	CT006	ge_57	461.7	-3.0	CZ018	ge_87	595.9	1.2
CT004	ge_27	472.4	-12.5	CT006	ge_58	586.7	-3.1	CZ018	ge_88	569.2	13.9
CT004	ge_28	450.6	-5.8	CT006	ge_59	325.0	0.3	CZ018	ge_89	969.5	1.0
CT004	ge_29	812.2	1.4	CT006	ge_60	537.8	-2.4	CZ018	ge_90	592.1	2.7
CT004	ge_30	768.6	7.6	CT006	ge_61	492.0	-8.5				

Table 2.5 Continued

Sample	Analysis	Age (Ma)	ϵ_{Hf}	Sample	Analysis	Age (Ma)	ϵ_{Hf}	Sample	Analysis	Age (Ma)	ϵ_{Hf}
CZ018	ge_91	961.7	0.8	CT011	de_44	492.9	-28.2	IG22-18	gf_1	94.9	9.3
CZ018	ge_92	1050.6	2.1	CT011	de_30	507.6	-13.5	IG22-18	gf_2	94.9	12.0
CZ018	ge_93	554.8	-7.8	CT011	de_40	513.3	-17.4	IG22-18	gf_3	94.9	12.1
CZ018	ge_94	966.9	6.2	CT011	de_28	536.9	-9.3	IG22-18	gf_4	94.9	10.7
CZ018	ge_95	987.8	3.6	CT011	de_38	539.6	-17.4	IG22-18	gf_5	94.9	12.4
CZ018	ge_96	558.3	1.4	CT011	de_37	541.0	-60.6	IG22-18	gf_6	94.9	11.3
CZ018	ge_97	524.9	3.2	CT011	de_45	541.0	-11.0	IG22-18	gf_7	94.9	10.4
IG22-43	gf_98	1292.3	-6.9	CT011	de_41	547.0	-31.8	IG22-18	gf_8	94.9	11.2
IG22-43	gf_99	1584.6	-6.0	CT011	de_43	581.4	-24.9	IG22-19	gf_9	91.4	13.3
IG22-43	gf_100	275.4	2.6	CT011	de_42	600.4	-11.4	IG22-19	gf_10	91.4	10.7
IG22-43	gf_101	269.4	-2.4	CT011	de_29	604.1	-22.2	IG22-19	gf_11	91.4	11.5
IG22-43	gf_102	265.5	1.6	CT011	de_47	681.9	-12.6	IG22-19	gf_12	91.4	8.5
IG22-43	gf_103	200.5	-2.1	CT011	de_33	686.5	-13.8	IG22-19	gf_13	91.4	8.6
IG22-43	gf_104	313.6	-4.1	CT011	de_36	692.9	-11.8	IG22-19	gf_14	91.4	13.1
IG22-10	gc_15	191.4	3.0	CT011	de_21	735.6	-12.6	IG22-51B	gf_64	101	12.1
IG22-10	gc_16	724.8	6.6	CT011	de_18	745.5	-24.2	IG22-51B	gf_65	101	11.8
IG22-10	gc_17	619.8	3.7	CT011	de_46	809.7	-22.1	IG22-51B	gf_66	101	11.0
IG22-10	gc_19	257.8	-14.4	CT011	de_17	931.8	-11.0	IG22-51B	gf_67	101	11.1
IG22-10	gc_20	732.3	-2.6	CT011	de_32	982.0	-11.6	IG22-51B	gf_68	101	11.8
IG22-10	gc_21	186.1	3.9	CT013	CT013_4	217.6	-2.6	IG22-51B	gf_69	101	11.3
IG22-09	gf_15	243.0	-7.3	CT013	CT013_5	218.9	-1.9	IG22-51B	gf_70	101	12.5
IG22-09	gf_16	231.0	-19.6	CT013	CT013_9	223.9	-1.3	IG22-28	jd6	138	6.0
IG22-09	gf_17	357.0	3.0	CT013	CT013_12	221.7	-6.9	IG22-28	jd7	138	4.8
IG22-09	gf_18	493.0	-3.2	CT013	CT013_13	224.6	-2.8	IG22-28	jd8	138	5.0
IG22-09	gf_19	518.0	-1.3	CT013	CT013_14	223.7	-2.5	IG22-28	jd9	138	7.0
IG22-09	gf_20	403.0	-3.4	CT013	CT013_15	223.8	-5.0	IG22-28	jd10	138	7.2
IG22-09	gf_21	499.0	16.3	CT013	CT013_16	224.0	-7.9	IG22-22	gf_22	186	5.8
CT009	de_4	374.9	-14.0	CT013	CT013_19	221.4	-8.3	IG22-22	gf_23	186	5.5
CT009	de_7	377.9	-39.0	CT013	CT013_18	222.9	-0.8	IG22-22	gf_24	186	5.5
CT009	de_9	469.9	-24.0	CT010	CT010_21	560.6	-17.3	IG22-22	gf_25	186	4.9
CT009	de_16	509.7	-10.6	CT010	CT010_22	516.3	-15.9	IG22-22	gf_26	186	5.4
CT009	de_11	523.1	-21.3	CT010	CT010_9	472.4	-16.0	IG22-22	gf_27	186	5.0
CT009	de_1	523.7	-25.5	CT010	CT010_8	557.9	-18.9	IG22-22	gf_28	186	4.5
CT009	de_14	532.5	-10.7	CT010	CT010_7	525.1	-16.3	IG22-23	gf_29	191	5.9
CT009	de_8	565.4	-32.1	CT010	CT010_4	1051.1	-28.8	IG22-23	gf_30	191	7.5
CT009	de_10	585.4	-11.4	CT010	CT010_19	342.1	-11.9	IG22-23	gf_31	191	5.5
CT009	de_15	624.5	-28.9	CT010	CT010_1	351.6	-7.8	IG22-23	gf_32	191	6.7
CT009	de_2	694.8	-15.1	CT010	CT010_63	1081.9	-17.8	IG22-23	gf_33	191	5.2
CT009	de_6	897.1	-26.9	CT010	CT010_49	1049.7	-18.5	IG22-23	gf_34	191	7.4
CT009	de_3	1038.7	-17.2	IG22-34	gf_71	187.8	5.1	IG22-23	gf_35	191	5.5
CT009	de_5	1045.2	-17.0	IG22-34	gf_72	187.8	6.6	IG22-26	jd86	179	4.2
CT011	de_25	340.3	-15.0	IG22-34	gf_73	187.8	5.9	IG22-26	jd87	179	6.5
CT011	de_39	347.4	-13.4	IG22-34	gf_74	187.8	5.8	IG22-26	jd88	179	0.0
CT011	de_23	348.2	-17.0	IG22-34	gf_75	187.8	5.3	IG22-26	jd90	179	5.1
CT011	de_31	488.2	-13.6	IG22-34	gf_77	187.8	6.1	IG22-53	gf_58	236	-1.4

Table 2.5 Continued

Sample	Analysis	Age (Ma)	ϵ_{Hf}	Sample	Analysis	Age (Ma)	ϵ_{Hf}	Sample	Analysis	Age (Ma)	ϵ_{Hf}
IG22-53	gf_59	235.5	0.1	IG22-31	jd19	133.7	8.1	SAN16	jc67	33.6	-0.8
IG22-53	gf_60	235.5	-2.6	IG22-31	jd20	133.7	7.8	SAN16	jc68	33.6	-1.3
IG22-53	gf_61	235.5	-3.6	SAN18	jc76	139.8	7.5	SAN16	jc69	33.6	2.2
IG22-53	gf_62	235.5	-3.2	SAN18	jc77	139.8	8.1	SAN16	jc70	33.6	0.3
IG22-53	gf_63	235.5	-11.0	SAN18	jc78	139.8	8.1	SAN14	jc11	33.7	0.8
IG22-47	ga_13	93.3	-1.9	SAN18	jc79	139.8	6.6	SAN14	jc12	33.7	0.9
IG22-47	ga_14	93.3	-10.1	SAN18	jc80	139.8	7.3	SAN14	jc13	33.7	0.3
IG22-47	ga_15	93.3	8.6	IG22-32	gf_85	157.1	5.2	SAN14	jc14	33.7	-0.8
IG22-48	gd_43	97.2	12.0	IG22-32	gf_86	157.1	5.3	SAN14	jc15	33.7	-0.9
IG22-48	gd_44	97.2	12.1	IG22-32	gf_87	157.1	4.5	SAN15	jc16	33.7	-0.1
IG22-48	gd_45	97.2	11.7	IG22-32	gf_89	157.1	6.5	SAN15	jc17	33.7	-0.9
IG22-48	gd_46	97.2	10.6	IG22-32	gf_90	157.1	5.7	SAN15	jc18	33.7	-1.9
IG22-48	gd_47	97.2	11.1	SAN19	jd1	19.9	3.5	SAN15	jc19	33.7	-0.2
IG22-48	gd_48	97.2	10.7	SAN19	jd2	19.9	2.9	SAN15	jc20	33.7	-1.1
IG22-48	gd_49	97.2	12.3	SAN19	jd3	19.9	4.8	SAN13	jc61	34.3	-1.0
IG22-49	gd_29	87.8	11.5	SAN19	jd4	19.9	3.5	SAN13	jc62	34.3	-0.3
IG22-49	gd_30	87.8	11.3	SAN19	jd5	19.9	2.6	SAN13	jc63	34.3	-0.3
IG22-49	gd_31	87.8	10.9	SAN08	jc41	20.2	4.5	SAN13	jc64	34.3	-1.5
IG22-49	gd_32	87.8	10.3	SAN08	jc42	20.2	3.8	SAN13	jc65	34.3	-0.5
IG22-49	gd_33	87.8	9.9	SAN08	jc43	20.2	5.3	lg22-33	gf_91	35.9	4.9
IG22-49	gd_34	87.8	10.0	SAN08	jc44	20.2	5.1	lg22-33	gf_92	35.9	5.8
IG22-49	gd_35	87.8	10.5	SAN08	jc45	20.2	3.9	lg22-33	gf_93	35.9	6.7
IG22-50	gc_8	97.6	7.0	SAN09	jc46	20.2	5.0	lg22-33	gf_94	35.9	5.4
IG22-50	gc_9	97.6	14.2	SAN09	jc47	20.2	4.0	lg22-33	gf_95	35.9	4.6
IG22-50	gc_10	97.6	10.4	SAN09	jc48	20.2	4.0	lg22-33	gf_96	35.9	4.8
IG22-50	gc_12	97.6	3.1	SAN09	jc49	20.2	4.2	lg22-33	gf_97	35.9	5.3
IG22-50	gc_13	97.6	11.4	SAN09	jc50	20.2	4.1	IG22-39	gf_50	40.7	5.0
IG22-50	gc_14	97.6	11.7	SAN10	jc51	29.1	4.6	IG22-39	gf_51	40.7	4.9
IG22-51°	jd81	99.1	13.8	SAN10	jc52	29.1	4.8	IG22-39	gf_52	40.7	2.4
IG22-51°	jd82	99.1	13.1	SAN10	jc53	29.1	5.0	IG22-39	gf_53	40.7	2.5
IG22-51°	jd83	99.1	13.6	SAN10	jc54	29.1	4.0	IG22-39	gf_55	40.7	3.6
IG22-51°	jd84	99.1	14.8	SAN10	jc55	29.1	3.6	IG22-39	gf_56	40.7	3.0
IG22-51°	jd85	99.1	13.3	IG22-38	gf_36	29.6	-3.9	SAN05	jc21	42.0	5.0
SAN01	jc6	90.0	11.3	IG22-38	gf_37	29.6	-2.7	SAN05	jc22	42.0	6.0
SAN01	jc7	90.0	12.2	IG22-38	gf_38	29.6	-3.3	SAN05	jc23	42.0	8.2
SAN01	jc8	90.0	10.3	IG22-38	gf_39	29.6	-3.1	SAN05	jc24	42.0	4.5
SAN01	jc9	90.0	12.1	IG22-38	gf_40	29.6	-2.8	SAN05	jc25	42.0	6.3
SAN01	jc10	90.0	10.4	IG22-38	gf_41	29.6	-2.3	SAN06	jc1	43.6	5.0
IG22-30	jd11	126.1	11.9	IG22-38	gf_42	29.6	-3.6	SAN06	jc2	43.6	4.5
IG22-30	jd12	126.1	7.3	IG22-35	gf_78	32.7	0.2	SAN06	jc3	43.6	5.3
IG22-30	jd13	126.1	10.6	IG22-35	gf_79	32.7	-1.5	SAN06	jc4	43.6	6.3
IG22-30	jd14	126.1	10.2	IG22-35	gf_80	32.7	0.6	SAN06	jc5	43.6	5.7
IG22-30	jd15	126.1	8.0	IG22-35	gf_81	32.7	0.3	SAN03	jc36	44.1	2.9

Table 2.5 continued

Sample	Analysis	Age (Ma)	εHf	Sample	Analysis	Age (Ma)	εHf	Sample	Analysis	Age (Ma)	εHf
SAN03	jc40	44.1	3.5	SAN11	jc56	58.9	3.1	SAN16	jc66	33.6	1.3
SAN02	jc31	45.2	3.9	SAN11	jc57	58.9	4.7	IG22-31	jd16	133.7	8.8
SAN02	jc32	45.2	3.2	SAN11	jc58	58.9	4.9	IG22-31	jd17	133.7	9.0
SAN02	jc33	45.2	2.8	SAN11	jc59	58.9	4.7	IG22-31	jd18	133.7	7.3
SAN02	jc34	45.2	3.5	SAN11	jc60	58.9	4.3	IG22-35	gf_82	32.7	-1.3
SAN02	jc35	45.2	4.0	SAN17	jc71	40.2	-1.4	IG22-35	gf_83	32.7	0.9
SAN04	jc26	46.7	3.8	SAN17	jc72	40.2	-2.8	IG22-35	gf_84	32.7	-1.6
SAN04	jc27	46.7	3.9	SAN17	jc73	40.2	-2.9	SAN03	jc37	44.1	2.4
SAN04	jc28	46.7	3.8	SAN17	jc74	40.2	-5.2	SAN03	jc38	44.1	2.1
SAN04	jc29	46.7	2.3	SAN17	jc75	40.2	-4.8	SAN03	jc39	44.1	1.6
SAN04	jc30	46.7	3.1	IG22-34	gf_76	187.8	5.5	CT011	de_35	352.1	-36.5
IG22-53	gf_57	236	1.7								

2.6.2 Oxygen $\delta^{18}\text{O}$ analyses

Oxygen exists in various isotopic forms, including oxygen-16, oxygen-17, and oxygen-18, which differ in their neutron counts. The delta (δ) notation is used to compare the ratio of ^{18}O to ^{16}O in a sample to that of a standardized reference material, typically Vienna Standard Mean Ocean Water (VSMOW). VSMOW represents a consistent isotopic composition of ocean water and serves as the reference standard (Kendall & Caldwell, 1998). This comparison is expressed using the following equation:

$$\delta^{18}\text{O} = \left[\frac{\left(\frac{^{18}\text{O}}{^{16}\text{O}} \right)_{\text{sample}}}{\left(\frac{^{18}\text{O}}{^{16}\text{O}} \right)_{\text{standard}}} - 1 \right] * 1000\text{‰} \quad \text{equation 2.4}$$

The analysis of $\delta^{18}\text{O}$ in igneous zircons with known ages is a powerful tool for tracing the evolution of intracrustal recycling and interactions between the crust and mantle over time. Oxygen isotope ratios in igneous rocks are significantly influenced by the incorporation of supracrustal materials, which typically exhibit higher $\delta^{18}\text{O}$ values compared to primitive mantle magmas. This makes $\delta^{18}\text{O}$ data particularly valuable for understanding the contributions of crustal materials to magmatic processes (Valley et al., 2005).

Oxygen isotopic compositions in zircon were measured using a Cameca IMS1270 at the Centre de Recherches Pétrographiques et Géochimiques (CRPG), Université de Lorraine, Nancy, France. The analyses employed a Cs^+ primary ion beam accelerated at 10 kV with an intensity of 2.5 nA. The detection mode utilized FC/FC multicollection. Instrumental mass fractionation was corrected using

the 91500-zircon reference material, for which the $\delta^{18}\text{O}$ value was determined to range from 4.41‰ to 6.42‰ via BrF_5 extraction and gas source mass spectrometry. Measured $^{18}\text{O}/^{16}\text{O}$ ratios were normalized to Vienna Standard Mean Ocean Water (VSMOW) compositions, with a reference ratio of $^{18}\text{O}/^{16}\text{O} = 0.0020052$ (Craig & Gordon, 1965). These ratios were further corrected for instrumental mass fractionation (IMF). A total of 160 new analyses are reported in Table 2.6 and displayed in Fig. 2.7.

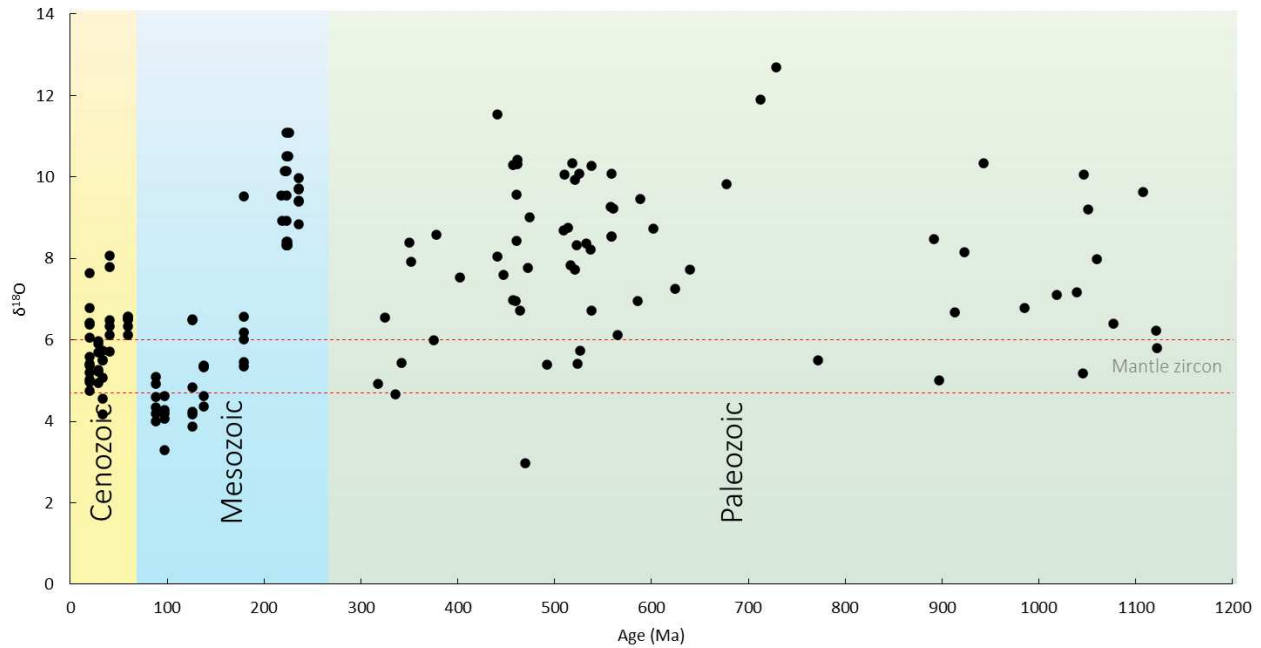


Figure 2.7 $\delta^{18}\text{O}$ vs. Age isotopic data on zircon

Table 2.6 zircon $\delta^{18}\text{O}$ data

Sample	Analysis	Age (Ma)	$\delta^{18}\text{O}$	Sample	Analysis	Age (Ma)	$\delta^{18}\text{O}$	Sample	Analysis	Age (Ma)	$\delta^{18}\text{O}$
SAN08	SAN-8@6	20.2	6.8	SAN16	SAN-16@5	33.6	4.2	SAN17	SAN-17@5	40.2	5.7
SAN10	SAN-10@1	29.1	5.7	SAN16	SAN-16@6	33.6	5.5	SAN17	SAN-17@6	40.2	6.3
SAN10	SAN-10@2	29.1	5.2	SAN17	SAN-17@1	40.2	6.5	SAN11	SAN-11@1	59	6.5
SAN10	SAN-10@3	29.1	5.3	SAN17	SAN-17@2	40.2	6.1	SAN11	SAN-11@2	59	6.3
SAN10	SAN-10@4	29.1	5.9	SAN17	SAN-17@3	40.2	7.8	SAN11	SAN-11@3	59	6.5
SAN10	SAN-10@5	29.1	4.9	SAN17	SAN-17@4	40.2	8.1	SAN11	SAN-11@4	59	6.1
SAN10	SAN-10@6	29.1	6.0	SAN11	SAN-11@6	59	6.6	SAN11	SAN-11@5	59	6.6
SAN16	SAN-16@1	33.6	5.7	IG22-53	IG22-53@6	235.5	9.7	CT010	CT10@49	1077.2	6.4
SAN16	SAN-16@2	33.6	5.5	SAN19	SAN-19@1	19.9	6.1	CT010	CT10@52	456.8	7.0
SAN16	SAN-16@3	33.6	4.6	SAN19	SAN-19@2	19.9	6.4	CT010	CT10@55	441.1	11.5
SAN16	SAN-16@4	33.6	5.1								

Table 2.6 Continued

Sample	Analysis	Age (Ma)	$\delta^{18}\text{O}$	Sample	Analysis	Age (Ma)	$\delta^{18}\text{O}$	Sample	Analysis	Age (Ma)	$\delta^{18}\text{O}$
IG22-49	IG22-49@1	87.8	4.6	CT013	CT013@13	224.6	10.5	CT010	CT10@53	440.9	8.1
IG22-49	IG22-49@2	87.8	4.4	CT013	CT013@14	225.6	11.1	CT010	CT10@54	513.7	8.8
IG22-49	IG22-49@3	87.8	4.0	CT013	CT013@15	223.8	8.3	CT010	CT10@60	639.7	7.7
IG22-49	IG22-49@4	87.8	4.2	CT013	CT013@16	224	8.3	CT010	CT10@59	460.9	9.6
IG22-49	IG22-49@5	87.8	5.1	CT013	CT013@19	221.4	10.1	CT010	CT10@66	520.8	9.9
IG22-49	IG22-49@6	87.8	4.9	CT013	CT013@9	223.9	8.4	CT006	CT06@1	509.4	8.7
IG22-50	IG22-50@1	97.6	4.2	CT013	CT013@4	217.6	9.5	CT006	CT06@02	537.4	8.2
IG22-50	IG22-50@2	97.6	3.3	CT013	CT013@5	218.9	8.9	CT006	CT06@03	1018.2	7.1
IG22-50	IG22-50@3	97.6	4.3	CT009	CT009G@49	897.1	5.0	CT006	CT06@04	474.1	9.0
IG22-50	IG22-50@4	97.6	4.6	CT009	CT009G@22	377.9	8.6	CT006	CT06@05	537.7	10.3
IG22-50	IG22-50@5	97.6	0.3	CT009	CT009G@51	565.4	6.1	CT006	CT06@06	464.2	6.7
IG22-50	IG22-50@6	97.6	4.1	CT009	CT009G@55	469.9	3.0	CT006	CT06@08	456.7	10.3
IG22-28	IG22-28@1	137.5	5.3	CT009	CT009G@25	585.4	7.0	CT006	CT06@09	891.3	8.5
IG22-28	IG22-28@2	137.5	5.3	CT009	CT009G@17	523.1	8.3	CT006	CT06@10	335.8	4.7
IG22-28	IG22-28@3	137.5	4.4	CT009	CT009G@13	1121.7	5.8	CT006	CT06@12	447.1	7.6
IG22-28	IG22-28@4	137.5	5.4	CT009	CT009G@12	469.5	-1.6	CT006	CT06@13	677.2	9.8
IG22-28	IG22-28@5	137.5	5.3	CT009	CT009G@11	985	6.8	CT006	CT06@14	317.2	4.9
IG22-28	IG22-28@6	137.5	4.6	CT009	CT009G@6	532.5	8.4	CT006	CT06@15	461	8.4
IG22-30	IG22-30@1	126.1	4.2	CT009	CT009G@4	624.5	7.3	CT006	CT06@16	461.5	10.4
IG22-30	IG22-30@2	126.1	4.8	CT009	CT009G@3	509.7	10.1	CT006	CT06@17	712.5	11.9
IG22-30	IG22-30@3	126.1	6.5	CT009	CT009G@29	1038.7	7.2	CT006	CT06@18	728.7	12.7
IG22-30	IG22-30@4	126.1	6.5	CT009	CT009G@27	523.7	5.4	CT006	CT06@21	588.5	9.5
IG22-30	IG22-30@5	126.1	3.9	CT009	CT009G@32	694.8	0.4	CT006	CT06@22	492	5.4
IG22-30	IG22-30@6	126.1	4.2	CT009	CT009G@38	374.9	6.0	CT006	CT06@24	526.3	5.7
IG22-26	IG22-26@1	179.2	9.5	CT009	CT009G@39	1045.2	5.2	CT006	CT06@25	942.8	10.3
IG22-26	IG22-26@2	179.2	5.5	CT010	CT10@68	521	7.7	CT006	CT06@26	325	6.6
IG22-26	IG22-26@3	179.2	6.0	CT010	CT10@63	1121	6.2	CT006	CT06@27	537.8	6.7
IG22-26	IG22-26@4	179.2	6.2	CT010	CT10@71	558.7	8.5	CT006	CT06@28	461.7	10.3
IG22-26	IG22-26@5	179.2	6.6	CT010	CT10@17	559.1	10.1	CT006	CT06@29	1059.5	8.0
IG22-26	IG22-26@6	179.2	5.3	CT010	CT10@18	402.3	7.5	CT006	CT06@31	459.8	7.0
CT013	CT013@13	222.8	10.5	CT010	CT10@71	558.7	8.5	CT006	CT06@32	923.2	8.2
CT013	CT013@14	222.8	11.1	CT010	CT10@73	602.2	8.7	CT006	CT06@34	518.3	10.3
CT013	CT013@15	222.8	8.3	CT010	CT10@01	351.6	7.9	CT006	CT06@35	913.2	6.7
CT013	CT013@16	222.8	8.3	CT010	CT10@02	1107.7	9.6	CT006	CT06@36	771.5	5.5
CT013	CT013@19	222.8	10.1	CT010	CT10@19	342.1	5.4	SAN19	SAN-19@3	19.9	5.4
CT013	CT013@9	222.8	8.4	CT010	CT10@20	349.6	8.4	SAN19	SAN-19@4	19.9	5.4
CT013	CT013@4	222.8	9.5	CT010	CT10@04	1046.4	10.1	SAN19	SAN-19@5	19.9	5.6
CT013	CT013@5	222.8	8.9	CT010	CT10@07	525.1	10.1	SAN19	SAN-19@6	19.9	5.0
IG22-53	IG22-53@1	235.5	9.7	CT010	CT10@8	557.9	9.3	SAN08	SAN-8@1	20.2	5.0
IG22-53	IG22-53@2	235.5	9.4	CT010	CT10@9	472.4	7.8	SAN08	SAN-8@2	20.2	5.2
IG22-53	IG22-53@3	235.5	9.4	CT010	CT10@21	560.6	9.2	SAN08	SAN-8@3	20.2	6.4
IG22-53	IG22-53@4	235.5	10.0	CT010	CT10@22	516.3	7.8	SAN08	SAN-8@4	20.2	7.6
IG22-53	IG22-53@5	235.5	8.8	CT010	CT10@36	1051.2	9.2	SAN08	SAN-8@5	20.2	4.8

2.6.3 Trace elements in zircon

The trace element (TE) composition of zircon has gained significant attention due to its invaluable role in unraveling the evolution of long-lived and complex silicic magmatic systems (Claiborne et al., 2010). Zircon grains can undergo multiple episodes of sedimentation, magmatism, and/or metamorphism while retaining both geochronological and geochemical information (Grimes et al., 2007). Consequently, the elemental composition of zircon provides critical insights into the development of magmatic processes and the evolution of tectonic settings (Belousova et al., 2002; Claiborne et al., 2010).

The analysis of zircon chemistry is challenging due to numerous influencing factors, including crystallization conditions, pressure, composition of coexisting minerals, and the nature of the magma source (Grimes et al., 2007; 2015). Despite these complexities, significant progress has been made in using TE data to interpret geodynamic contexts (Belousova et al., 2002; Grimes et al., 2015; Wang et al., 2012). Zircon's strong affinity for uranium (U), ytterbium (Yb) and other heavy rare earth elements (HREEs) allows these elements to be incorporated into its crystal structure. As a result, several plots such as U versus Yb, U/Yb ratio versus Hf and Y, as well as Th versus Yb and Th/Yb ratio versus Hf, have proven to be effective in distinguishing between various tectonic settings (Grimes et al., 2007).

Zircon trace-element concentrations were measured at Géosciences Montpellier (University of Montpellier, AETE-ISO regional facility of the OSU OREME) using laser ablation inductively coupled plasma mass spectrometry (LA-ICP-MS). The analytical platform comprised a 193 nm ArF excimer laser (Teledyne G2) coupled to a Thermofinnigan Element XR mass spectrometer, which was enhanced with an 80 m³/h Edwards primary pump at the interface to improve sensitivity. Measurements were conducted during two sessions, each with different analytical conditions.

In the first session, the laser was operated at a repetition rate of 6 Hz, using a 50 µm spot size and a fluence of 4.5 J/m². In the second session, the laser was operated at a repetition rate of 3 Hz, with a 26 µm spot size and an energy density of 12 J/cm². For both sessions, the total analysis time was 120 s, with the first 80 s allocated for background measurement before sample ablation. External calibration was performed using NIST 612, with reference values from Pearce et al. (2007). To ensure accuracy and long-term reproducibility, three zircon standards were analyzed in both sessions: 91500 (n = 20 and n = 53, respectively), GJ1 (n = 11 and n = 48), and Plešovice (n = 11 and n = 48).

Data reduction was performed using the Glitter software package (van Achterberg et al., 2001). We present 373 new measurements, with selected elemental data summarized in Table 2.7 for conciseness

and improved presentation. Detailed datasets are provided in subsequent chapters for comprehensive analysis.

Table 2.7 Zircon trace element data

Sample	Analysis	Age	La	Pr	Sm	Eu	Gd	Dy	Yb	Lu	Hf	U
CT004	ge_30	768.6	0.5050	1.18	18.02	4.30	54.8	218.1	536.4	81.3	8284.9	325.2
CT004	ge_31	494.9	0.0245	0.21	8.74	1.35	37.5	208.9	651.7	96.5	9813.0	140.7
CT004	ge_32	487.6	0.0232	0.04	2.90	0.32	19.7	202.2	1083.0	174.3	11310.6	294.3
CT004	ge_33	942.3	0.0692	0.29	5.78	1.16	22.0	113.9	367.7	54.0	10183.8	381.8
CT004	ge_34	1081.7	0.0109	0.10	4.19	0.57	17.4	90.5	301.6	46.6	9374.3	209.1
CT004	ge_35	519.1	0.0209	0.25	4.62	0.56	6.3	11.6	20.7	2.9	10782.0	188.8
CT004	ge_36	465.9	0.0287	0.06	2.32	0.44	8.6	42.3	200.2	34.0	9980.9	204.8
CT004	ge_37	906.9	0.0820	0.16	6.06	0.90	28.8	167.8	482.8	71.6	10720.0	301.6
CT004	ge_39	738.9	0.0048	0.04	2.44	0.17	13.7	97.2	381.4	63.4	11743.9	380.6
CT004	ge_40	802.4	0.0324	0.27	12.12	0.37	61.4	397.0	1003.9	130.6	9400.9	362.0
CT004	ge_41	997.4	0.0247	0.13	6.65	1.45	34.2	199.6	716.3	113.2	9990.7	369.9
CT004	ge_42	400.9	0.2390	0.22	4.27	2.04	17.9	92.2	432.2	67.0	9619.5	487.8
CT004	ge_43	526.0	0.1170	0.21	4.77	0.55	29.3	193.2	435.3	67.0	13804.6	759.7
CT004	CT04-30	319.2	0.0540	0.22	8.57	1.55	33.7	118.9	289.0	53.0	10040.0	182.7
CT004	CT4-29	485.4	0.0344	0.28	8.63	3.22	39.4	143.4	517.0	106.1	9490.0	347.0
CT004	CT04-20	1712.1	0.0730	0.17	1.67	0.60	6.8	39.5	289.3	66.3	14030.0	392.0
CT004	CT04-18	1437.2	0.4400	0.48	7.03	2.63	31.7	105.8	272.0	51.6	8090.0	252.9
CT004	CT04-14	1040.6	0.2310	0.12	2.61	0.24	16.0	71.1	238.1	47.8	9640.0	58.7
CT004	CT04-9	635.3	11.4000	3.60	6.40	1.18	17.5	73.6	302.0	59.4	8710.0	164.6
CT004	CT04-8	1988.9	13.8000	7.40	21.00	1.04	64.0	224.0	510.0	85.0	9910.0	114.8
CT004	CT04-11	785.7	0.0440	0.59	11.17	2.42	39.9	124.5	305.9	54.1	9910.0	183.1
CT004	CT04-12	1050.4	0.0290	0.23	5.72	0.82	19.9	65.8	232.2	43.5	13850.0	290.6
CT004	CT04-6	960.0	0.0000	0.12	4.71	0.37	26.5	107.1	293.6	53.2	11450.0	111.4
CT004	CT04-4	372.6	0.2160	1.16	18.10	4.44	59.5	162.6	319.0	53.4	9980.0	137.9
CT004	CT04-1	1126.2	23.6000	6.64	14.40	1.57	29.3	73.5	185.1	34.3	13340.0	124.6
CT006	ge_44	474.1	0.0842	0.19	4.06	0.65	22.4	231.5	1316.3	203.2	12747.9	544.7
CT006	ge_45	464.2	0.0331	0.17	15.01	6.07	59.9	142.3	396.2	69.5	10855.4	228.5
CT006	ge_46	461.0	0.0396	0.09	2.90	0.70	14.7	89.6	514.2	89.8	9464.6	85.8
CT006	ge_47	317.2	0.0565	0.17	6.06	1.75	28.3	153.6	602.4	103.6	10547.0	327.4
CT006	ge_48	447.1	0.0980	0.21	6.85	1.25	39.9	280.9	1091.2	196.3	11487.9	188.7
CT006	ge_49	461.5	0.0587	0.19	6.58	1.41	34.8	223.8	975.4	158.3	11843.1	283.0
CT006	ge_50	517.6	0.0715	0.22	6.89	1.61	34.3	169.2	649.3	103.1	9728.9	251.6
CT006	ge_52	518.3	0.0252	0.21	7.16	0.65	34.5	187.2	661.9	100.0	10064.1	210.4
CT006	ge_53	477.2	0.0103	0.02	1.51	0.23	8.5	51.0	243.1	40.8	9587.3	121.4
CT006	ge_54	769.3	0.0189	0.11	4.76	1.19	19.2	78.3	190.8	29.8	7450.6	66.5
CT006	ge_55	459.8	0.1320	0.11	2.43	0.65	11.1	54.6	245.6	41.7	10406.8	161.5
CT006	ge_56	330.4	0.0185	0.12	3.90	0.92	18.7	100.3	475.0	72.0	9531.3	270.3
CT006	ge_57	461.7	0.0572	0.05	2.37	0.25	18.4	192.1	995.1	161.7	11603.2	235.7
CT006	ge_58	586.7	0.0083	0.04	2.67	0.40	12.5	87.0	539.3	81.2	12775.4	382.7
CT006	ge_59	325.0	0.0405	0.24	8.60	2.50	42.2	209.1	848.2	136.6	9655.2	316.2

Table 2.7 continued

Sample	Analysis	Age	La	Pr	Sm	Eu	Gd	Dy	Yb	Lu	Hf	U
CT006	ge_60	537.8	0.2650	0.83	15.98	4.45	47.9	201.3	777.8	120.4	10003.4	487.3
CT006	ge_61	492.0	0.0612	0.22	6.87	1.11	28.9	116.0	287.0	41.3	8472.1	113.8
CT006	ge_62	456.7	0.0090	0.05	2.98	0.84	17.7	109.3	598.3	102.8	9902.0	288.6
CT006	ge_63	449.2	0.0268	0.05	1.24	0.50	6.1	30.5	188.8	27.8	12309.8	309.6
CT009	ge_64	223.8	0.0106	0.16	7.96	0.54	45.6	217.1	614.6	91.9	9114.1	334.7
CT009	ge_65	307.0	0.0062	0.06	2.78	0.14	17.5	103.4	554.3	84.5	11386.4	340.5
CT009	ge_66	779.0	0.0461	0.37	9.87	1.65	39.0	138.1	350.6	52.9	8807.1	55.7
CT009	ge_67	569.4	0.1480	1.05	21.81	4.22	77.7	234.8	449.9	63.7	8626.6	267.4
CT009	ge_68	463.2	0.0306	0.12	7.14	0.33	37.5	165.8	366.5	69.0	10082.1	256.5
CT009	ge_69	491.1	0.0103	0.04	1.98	0.15	9.3	37.0	106.6	16.3	10595.5	57.9
CT009	ge_70	1026.6	0.0175	0.18	9.66	0.17	59.4	297.8	883.4	133.4	11646.8	332.8
CT009	ge_71	528.6	0.0113	0.05	2.20	0.73	10.7	57.5	302.7	52.2	10930.7	149.7
CT009	ge_72	469.5	0.0528	0.08	3.93	0.59	24.0	133.8	464.2	100.6	9858.7	456.9
CT009	ge_73	1050.5	0.0375	0.19	7.42	0.85	36.8	149.0	407.9	60.1	9676.4	195.8
CT009	ge_74	472.7	0.0515	0.13	4.81	0.53	26.6	171.5	691.3	132.3	12048.4	488.1
CT009	ge_75	471.2	0.0084	0.05	3.62	0.97	22.0	120.2	437.9	89.9	12078.9	259.2
CT009	ge_76	569.3	0.0820	0.16	5.55	0.49	30.2	159.1	577.6	80.0	9637.9	249.7
CT009	ge_77	336.3	0.0216	0.10	5.03	0.24	32.7	165.5	525.5	73.8	10633.6	259.0
CT009	ge_78	495.0	0.1560	0.45	10.16	0.82	36.5	115.9	256.5	35.7	9296.9	118.8
CT010	ge_1	350.6	76.2900	24.59	29.33	1.39	40.7	156.8	497.5	75.8	9664.2	736.8
CT010	ge_2	442.6	0.0279	0.07	2.71	0.16	11.2	74.5	266.0	39.1	9778.7	378.1
CT010	ge_3	455.3	0.0713	0.09	4.92	0.47	22.1	155.1	560.6	80.9	10720.6	173.1
CT010	ge_4	356.0	0.0422	0.21	7.57	1.67	24.3	122.9	368.2	58.9	7447.2	129.5
CT010	ge_5	949.3	0.0623	0.17	3.91	0.77	13.7	75.8	234.3	37.0	8208.8	103.8
CT010	ge_6	525.7	0.0462	0.15	5.80	0.49	24.5	153.5	484.9	74.2	9818.0	124.9
CT010	ge_7	559.7	0.0126	0.01	0.35	0.17	1.6	7.3	11.2	1.6	11400.1	333.3
CT010	ge_8	424.3	0.2440	0.49	6.62	2.48	22.6	137.5	545.4	90.4	10455.2	551.6
CT010	ge_9	560.7	0.0222	0.06	3.02	0.29	11.0	59.3	196.5	30.0	10051.6	219.1
CT010	ge_10	402.3	0.5790	0.30	5.87	0.65	23.8	129.1	411.5	63.2	9723.1	339.8
CT010	ge_11	1069.0	0.1660	0.26	5.89	0.65	23.0	135.7	369.5	61.6	10425.5	282.4
CT010	ge_12	331.2	23.8400	11.33	38.76	4.30	92.7	406.2	1110.0	167.5	8479.1	291.2
CT010	ge_13	526.6	0.0065	0.07	4.17	0.34	21.2	162.6	639.3	99.2	10975.7	253.6
CT010	ge_14	525.2	0.0257	0.11	5.90	0.68	29.1	197.1	645.2	100.0	10417.8	131.7
CT010	ge_15	360.0	18.3500	7.13	15.65	1.82	35.9	158.4	442.9	69.9	9647.5	167.1
CT010	ge_16	329.8	0.0690	0.08	3.17	0.19	19.6	134.9	518.6	84.0	10408.4	425.3
CT010	ge_17	510.9	0.0148	0.06	3.05	0.74	13.9	93.5	429.4	74.9	9409.9	232.7
CT010	ge_18	352.0	0.4790	0.54	9.32	1.02	29.1	143.5	417.4	56.0	9560.2	175.6
CT010	ge_19	528.7	0.3100	0.35	5.27	0.63	23.2	158.6	506.0	74.7	12063.1	393.5
CT010	ge_20	559.5	0.0216	0.01	0.35	0.05	2.0	20.7	150.9	27.7	12890.2	221.7
CT010	ge_21	441.9	0.1780	0.30	6.80	1.84	27.5	156.4	643.6	102.7	10287.2	338.8
CT010	ge_22	527.3	0.0236	0.25	10.10	1.57	42.1	225.2	658.3	101.7	9191.5	190.1
CT010	ge_23	557.3	0.0912	0.12	4.66	0.81	22.5	128.2	439.8	70.4	9893.9	194.7
CT010	ge_24	1078.7	4.7100	2.33	15.15	1.81	48.6	216.7	514.4	83.0	8809.8	112.0
CT010	ge_25	560.2	0.1590	1.25	17.74	2.72	40.9	129.6	239.8	36.0	9088.9	141.9
CT010	ge_26	520.7	0.0165	0.11	6.50	0.43	32.3	191.0	614.5	94.9	11008.3	189.3
CT010	CT10-59	460.9	3.3700	1.07	8.20	0.86	35.9	120.5	228.0	56.5	10530.0	119.0

Table 2.7 continued

Sample	Analysis	Age	La	Pr	Sm	Eu	Gd	Dy	Yb	Lu	Hf	U
CT010	CT10-66	520.8	0.0087	0.20	9.24	1.43	57.3	235.0	562.0	106.6	11020.0	176.0
CT010	CT10-68	521.0	0.0012	0.12	5.89	1.11	28.7	108.9	290.0	57.5	9410.0	75.8
CT010	CT10-63	1081.9	0.0370	0.61	15.00	1.62	69.9	274.6	697.0	131.3	10460.0	187.4
CT010	CT10-17	559.1	0.0221	0.06	3.04	0.58	18.9	93.6	362.0	72.3	12830.0	208.0
CT010	CT10-18	402.3	0.0099	0.16	7.64	0.85	39.5	170.0	505.0	95.2	11770.0	506.0
CT010	CT10-71	558.7	0.0074	0.11	5.54	0.49	29.7	128.3	395.0	74.7	13520.0	224.0
CT010	CT10-73	602.2	0.0015	0.04	1.64	0.53	9.0	41.7	224.3	48.3	13240.0	485.0
CT010	CT10-2	1083.5	0.1080	0.14	2.94	0.57	15.1	68.8	298.3	62.8	11640.0	178.8
CT010	CT10-1	351.6	1.0400	0.35	4.88	1.48	23.8	100.5	373.0	75.7	11670.0	374.0
CT010	CT10-19	342.1	2.5900	1.05	6.29	0.83	33.2	181.3	717.0	141.2	12550.0	707.0
CT010	CT10-20	349.6	0.0890	0.22	8.17	0.37	44.7	184.1	521.0	98.8	12540.0	183.7
CT010	CT10-4	1051.1	0.0281	0.32	10.79	0.56	56.1	214.7	481.0	87.1	11750.0	137.2
CT010	CT10-8	557.9	0.0019	0.01	1.09	0.19	6.8	38.3	153.4	30.5	13510.0	79.2
CT010	CT10-9	472.4	2.4300	0.79	2.71	0.55	13.9	76.3	322.4	86.6	14680.0	374.9
CT010	CT10-7	525.1	0.0500	0.12	3.89	1.11	21.5	116.5	489.0	99.5	14340.0	436.0
CT010	CT10-21	560.6	0.0316	0.18	7.81	0.41	39.9	172.0	449.0	83.3	13500.0	312.0
CT010	CT10-22	516.3	0.0132	0.11	5.48	1.79	34.2	184.9	795.0	167.2	13350.0	475.0
CT010	CT10-36	1091.5	1.8900	1.28	13.17	0.70	56.6	208.4	475.0	86.4	12330.0	215.4
CT010	CT10-54	513.7	1.9300	0.81	3.22	0.51	17.5	91.2	357.0	69.7	13000.0	231.0
CT010	CT10-53	440.9	0.4400	0.23	2.46	0.54	10.4	52.0	257.7	53.9	12520.0	359.2
CT010	CT10-55	441.1	1.6000	0.53	6.38	0.37	37.9	287.8	1086.0	201.5	13500.0	337.0
CT010	CT10-52	456.8	0.0101	0.22	6.92	1.50	36.2	177.7	562.0	110.6	10740.0	131.8
CT010	CT10-49	1049.7	1.0600	0.30	4.02	0.20	22.6	134.3	486.0	93.9	13790.0	351.0
CT010	CT10-45	604.1	0.0950	0.21	4.19	0.36	15.3	56.5	148.7	28.9	10420.0	880.0
CT013	CT13-16	224.0	0.0570	0.81	28.40	1.71	154.0	593.0	1470.0	252.0	10970.0	571.0
CT013	CT13-19	221.4	0.0410	0.26	9.22	2.61	46.0	176.0	613.0	136.8	10380.0	301.7
CT013	CT13-15	223.8	18.1000	11.92	66.50	14.46	231.0	864.0	1632.0	302.0	12280.0	3800.0
CT013	CT13-14	223.7	17.2000	7.60	17.60	2.67	53.4	212.9	609.0	121.2	12390.0	1354.0
CT013	CT13-13	224.6	0.0410	0.40	20.70	2.63	131.0	585.0	1510.0	291.0	10100.0	1630.0
CT013	CT13-12	221.7	0.0410	0.45	17.65	2.82	87.1	313.0	713.0	142.8	11440.0	419.0
CT013	CT13-9	223.9	0.0600	0.10	4.95	0.71	33.0	164.9	517.0	107.7	10970.0	393.0
CT013	CT13-8	211.5	0.0064	0.16	6.65	0.72	36.9	150.0	402.0	75.2	9340.0	155.7
CT013	CT13-3	222.7	0.9320	2.42	25.70	6.13	89.8	278.0	634.0	117.4	9070.0	683.0
CT013	CT13-4	217.6	1.0900	2.66	41.50	16.30	115.0	278.0	507.0	87.9	11790.0	529.0
CZ018	ge_79	371.1	0.0260	0.28	11.53	0.76	54.1	208.4	565.3	84.6	9344.4	165.4
CZ018	ge_80	222.3	0.0545	0.22	9.17	0.52	49.8	217.2	662.1	102.5	8447.3	148.7
CZ018	ge_81	288.4	0.1120	0.14	6.18	0.11	52.2	382.5	1760.4	305.5	12580.4	614.2
CZ018	ge_82	361.9	0.0066	0.03	1.51	0.64	9.2	53.0	331.2	58.7	10205.0	442.6
CZ018	ge_83	241.5	0.0460	0.10	3.47	1.21	17.8	79.1	440.4	82.5	9986.4	175.7
CZ018	ge_84	452.1	0.0135	0.14	7.60	0.81	20.3	27.7	20.7	2.5	11874.9	555.9
CZ018	ge_85	420.5	0.0220	0.06	4.80	0.35	34.7	199.7	754.3	115.3	12016.7	222.4
CZ018	ge_86	595.9	0.2460	0.45	10.04	2.22	39.5	138.0	420.8	59.1	8307.1	100.4
CZ018	ge_87	569.2	0.0106	0.03	1.08	0.50	5.8	26.2	110.7	18.9	10706.4	85.0
CZ018	ge_88	969.5	0.0081	0.05	2.78	0.55	18.1	101.7	518.8	86.1	8338.1	129.4
CZ018	ge_89	592.1	0.0078	0.04	2.67	0.24	18.7	94.7	361.2	56.4	10897.9	194.3
CZ018	ge_90	961.7	0.0311	0.06	1.68	0.23	10.4	54.9	299.1	48.0	14404.8	205.0

Table 2.7 continued

Sample	Analysis	Age	La	Pr	Sm	Eu	Gd	Dy	Yb	Lu	Hf	U
CZ018	ge_91	1050.6	0.0348	0.22	7.51	1.69	33.4	118.5	313.5	44.5	8414.0	74.9
CZ018	ge_92	554.8	0.0014	0.03	1.47	0.19	9.9	42.8	175.0	26.6	9109.7	111.4
CZ018	ge_93	966.9	0.0156	0.17	8.43	0.20	30.0	54.2	59.9	7.5	11731.3	214.6
CZ018	ge_94	987.8	0.0209	0.20	6.54	1.16	28.9	88.1	245.8	38.1	6499.1	34.6
CZ018	ge_95	558.3	0.0132	0.21	12.29	0.44	73.7	255.8	443.6	59.3	11621.5	309.2
CZ018	ge_96	524.9	0.0092	0.03	1.82	0.12	13.9	81.3	446.5	67.2	13496.8	535.4
CZ018	ge_97	571.9	0.0361	0.21	4.58	1.85	19.9	77.6	402.3	72.7	8700.8	93.7
IG22-18	gf_9	94.9	11.5500	1.78	8.97	3.76	30.9	154.1	808.0	162.5	8045.1	4135.0
IG22-18	gf_10	94.9	0.4130	0.28	8.04	2.78	31.5	163.3	961.2	193.7	9097.7	5635.3
IG22-18	gf_11	94.9	0.6970	0.74	13.09	5.50	44.7	211.4	1144.2	222.2	8236.4	8129.9
IG22-18	gf_13	94.9	0.0230	0.23	6.67	1.89	24.0	103.6	323.8	53.6	9297.0	105.8
IG22-18	gf_14	94.9	1.7000	0.65	8.90	4.36	34.3	179.6	1168.9	233.2	6672.1	3925.2
IG22-19	gf_1	91.4	0.0079	0.02	0.79	0.44	3.4	25.5	195.8	37.7	7558.2	289.2
IG22-19	gf_2	91.4	0.0219	0.08	2.18	1.28	9.4	66.4	494.8	97.4	6767.6	787.2
IG22-19	gf_3	91.4	0.0309	0.08	2.32	1.37	11.2	82.5	578.0	116.6	7887.9	1144.3
IG22-19	gf_4	91.4	0.0590	0.07	1.71	1.03	7.0	46.7	386.3	76.3	9459.0	1067.5
IG22-19	gf_5	91.4	0.0109	0.03	1.04	0.59	5.9	51.6	455.3	98.8	9376.4	1121.7
IG22-19	gf_6	91.4	0.0124	0.02	0.93	0.62	6.3	55.1	469.3	100.0	9242.5	1260.9
IG22-19	gf_7	91.4	0.0086	0.02	0.82	0.46	4.2	32.4	272.5	56.4	7724.3	437.7
IG22-19	gf_8	91.4	0.0490	0.38	8.30	4.03	30.6	176.5	1276.9	252.9	6676.8	3841.4
IG22-22	gf_22	186.1	0.0069	0.05	1.56	0.67	7.6	48.6	366.5	74.5	9064.2	256.0
IG22-22	gf_23	186.1	0.0285	0.06	1.95	0.64	7.3	39.5	253.3	50.5	9432.9	125.7
IG22-22	gf_24	186.1	0.0060	0.02	0.68	0.33	4.0	27.8	235.3	49.6	9559.6	143.5
IG22-22	gf_25	186.1	0.0106	0.04	1.87	0.75	8.2	52.7	408.1	81.3	9381.7	271.6
IG22-22	gf_26	186.1	0.0365	0.20	6.18	1.95	24.1	127.9	732.9	131.8	9481.0	389.9
IG22-22	gf_27	186.1	0.0414	0.04	1.04	0.49	5.9	36.4	283.8	58.0	9468.2	216.1
IG22-22	gf_28	186.1	0.2660	0.10	0.96	0.55	6.5	44.2	315.5	69.2	9453.9	242.8
IG22-23	gf_29	191.4	0.0188	0.24	8.20	1.21	32.5	161.0	583.1	95.9	9016.8	186.5
IG22-23	gf_30	191.4	0.0075	0.12	4.90	0.87	22.5	126.7	497.3	83.2	9713.4	200.7
IG22-23	gf_31	191.4	0.0058	0.05	2.36	0.43	12.3	74.0	349.7	62.8	10287.0	175.1
IG22-23	gf_32	191.4	0.0899	0.09	3.35	0.65	17.4	92.6	360.1	59.7	9528.7	139.7
IG22-23	gf_33	191.4	0.0234	0.20	7.52	1.30	37.4	202.7	746.8	124.1	9218.1	361.1
IG22-23	gf_34	191.4	0.0336	0.21	7.14	1.32	35.6	186.7	707.5	117.2	9078.4	323.2
IG22-23	gf_35	191.4	0.0106	0.08	3.23	0.62	18.0	91.8	398.6	72.3	16836.5	558.8
IG22-26	jd_86	179.2	0.0180	0.14	5.02	1.19	28.9	115.4	450.5	77.2	10446.4	289.5
IG22-26	jd_87	179.2	0.0116	0.06	2.37	0.48	13.7	55.5	257.3	48.9	10486.4	157.8
IG22-26	jd_88	179.2	0.0021	0.02	1.39	0.50	8.4	31.4	147.7	27.4	11028.8	42.1
IG22-26	jd_89	179.2	0.0388	0.06	2.44	0.10	13.8	71.5	253.5	41.3	10018.4	231.7
IG22-26	jd_90	179.2	0.5780	0.22	1.96	0.33	10.2	43.8	218.6	43.1	9344.9	200.9
IG22-28	jd_6	137.5	0.0045	0.13	5.96	1.14	26.4	116.6	372.0	61.9	9199.2	108.3
IG22-28	jd_7	137.5	0.0024	0.04	1.76	0.40	8.7	46.8	192.1	34.3	9229.6	59.4
IG22-28	jd_8	137.5	0.0085	0.12	4.71	0.81	21.7	100.7	338.6	58.8	9125.5	96.2
IG22-28	jd_9	137.5	0.0014	0.03	1.46	0.31	8.0	44.8	210.2	38.1	10136.9	74.6
IG22-30	jd_11	126.1	0.0736	0.65	24.39	3.91	95.8	392.0	1167.1	191.0	9572.9	916.0
IG22-30	jd_12	126.1	0.0162	0.11	4.27	0.88	19.5	97.9	468.4	86.8	9614.3	404.2
IG22-30	jd_13	126.1	0.0377	0.38	12.24	1.79	48.3	195.8	594.1	101.3	9116.5	444.1

Table 2.7 continued

Sample	Analysis	Age	La	Pr	Sm	Eu	Gd	Dy	Yb	Lu	Hf	U
IG22-30	jd_14	126.1	1.2790	0.67	9.16	1.70	34.4	146.7	510.3	87.3	9820.3	337.8
IG22-30	jd_15	126.1	0.0220	0.16	4.81	0.90	21.0	92.5	342.0	62.1	8566.7	240.4
IG22-31	jd_16	133.7	0.0019	0.03	1.42	0.51	7.2	38.3	267.7	56.9	9116.4	240.1
IG22-31	jd_17	133.7	0.0074	0.07	2.87	0.90	11.9	56.6	320.2	66.7	9207.7	244.3
IG22-31	jd_18	133.7	0.0056	0.03	1.17	0.45	6.3	33.0	235.8	49.9	10402.8	373.2
IG22-31	jd_19	133.7	0.0028	0.03	1.16	0.44	5.6	29.2	189.4	40.0	9361.1	191.6
IG22-31	jd_20	133.7	0.0038	0.03	0.98	0.47	4.8	27.2	215.0	45.5	9201.7	204.8
IG22-32	gf_85	157.1	0.0154	0.07	1.62	0.76	8.0	32.9	211.5	43.1	8828.2	163.2
IG22-32	gf_86	157.1	0.0183	0.07	1.62	0.83	8.7	44.0	345.4	70.5	8441.1	351.5
IG22-32	gf_87	157.1	0.0316	0.21	4.69	1.98	25.7	107.4	553.1	103.1	7974.2	380.1
IG22-32	gf_89	157.1	0.0243	0.04	1.30	0.65	7.7	39.6	300.9	57.6	8621.0	295.7
IG22-32	gf_90	157.1	0.0155	0.12	2.70	1.29	13.5	53.0	284.4	53.9	8533.0	232.0
lg22-33	gf_91	35.9	0.0224	0.04	2.74	0.34	24.5	163.4	914.4	161.4	12574.6	499.8
lg22-33	gf_92	35.9	0.0126	0.07	4.76	0.80	31.4	176.1	833.8	134.8	10657.0	396.8
lg22-33	gf_93	35.9	0.0850	0.48	17.79	5.41	92.9	405.7	1982.7	364.2	7545.9	1199.8
lg22-33	gf_94	35.9	0.0050	0.02	2.27	0.40	19.0	120.4	778.2	135.0	10477.7	494.0
lg22-33	gf_95	35.9	1.4900	0.58	3.91	0.58	30.2	183.3	1136.1	185.4	11326.6	870.0
lg22-33	gf_96	35.9	0.0061	0.02	2.29	0.38	17.7	107.4	650.7	112.4	11152.6	464.6
lg22-33	gf_97	35.9	0.0506	0.15	7.30	1.42	40.7	182.3	819.4	129.2	8909.7	334.0
IG22-34	gf_71	187.8	0.0034	0.04	2.95	0.42	19.3	101.0	438.2	73.2	11025.2	267.4
IG22-34	gf_72	187.8	0.0295	0.29	10.34	1.37	53.0	232.9	817.8	133.0	9469.5	370.6
IG22-34	gf_73	187.8	0.0171	0.18	8.21	1.03	41.1	181.6	664.0	109.4	10120.5	321.2
IG22-34	gf_74	187.8	0.0098	0.08	4.37	0.58	24.5	119.6	505.0	82.4	10719.4	259.4
IG22-34	gf_75	187.8	0.0087	0.07	4.63	0.66	28.5	137.6	559.5	87.5	10775.1	242.7
IG22-34	gf_76	187.8	0.0244	0.26	9.50	1.28	53.1	237.1	841.9	141.6	10073.8	373.6
IG22-34	gf_77	187.8	0.0031	0.03	2.53	0.36	17.0	83.5	360.6	60.4	11286.6	261.4
IG22-35	gf_78	32.7	0.0297	0.35	13.98	2.51	75.3	287.8	787.6	123.0	8883.1	214.9
IG22-35	gf_79	32.7	0.0097	0.11	5.79	1.18	32.3	130.6	383.8	62.5	9740.5	86.4
IG22-35	gf_80	32.7	0.0109	0.13	5.35	0.98	29.6	127.1	423.8	66.8	9061.2	109.5
IG22-35	gf_81	32.7	0.0127	0.08	3.76	0.68	20.4	87.0	293.0	46.7	9535.3	76.5
IG22-35	gf_82	32.7	0.0187	0.22	8.61	1.53	47.9	192.0	561.3	87.8	9680.5	152.5
IG22-35	gf_83	32.7	0.0144	0.16	7.98	1.30	42.3	165.4	476.5	72.0	9903.2	127.9
IG22-35	gf_84	32.7	0.0094	0.16	6.36	1.19	34.2	136.0	421.3	65.6	9535.6	101.2
IG22-38	gf_36	29.6	0.0063	0.03	1.54	0.59	6.8	29.2	122.3	22.3	9506.6	26.8
IG22-38	gf_37	29.6	0.0058	0.03	1.54	0.53	7.0	33.0	145.8	26.5	10857.8	42.3
IG22-38	gf_38	29.6	0.0164	0.05	1.98	0.71	10.3	50.8	216.8	38.4	10049.0	57.1
IG22-38	gf_39	29.6	0.0124	0.09	3.77	1.40	16.1	73.4	280.4	48.8	8787.4	46.3
IG22-38	gf_40	29.6	0.0124	0.19	5.82	1.85	22.3	83.0	244.8	41.6	9463.4	58.1
IG22-38	gf_41	29.6	0.0102	0.08	2.93	1.06	12.7	52.4	204.8	37.0	9637.1	41.5
IG22-38	gf_42	29.6	0.0034	0.03	1.51	0.61	7.6	32.8	122.3	22.9	8993.6	20.0
IG22-39	gf_50	41.3	0.0085	0.08	2.80	0.79	15.3	86.5	550.1	104.3	9556.0	251.0
IG22-39	gf_51	41.3	0.0129	0.09	3.74	0.81	20.7	115.1	547.9	96.8	9657.0	177.9
IG22-39	gf_52	41.3	0.0054	0.05	4.22	0.88	30.0	159.4	600.7	101.2	8945.7	303.7
IG22-39	gf_55	41.3	0.0163	0.16	8.45	1.44	49.6	256.3	866.3	141.3	9421.1	234.7
IG22-39	gf_56	41.3	0.0134	0.11	5.63	1.01	29.8	161.0	606.8	100.4	9558.7	156.8
IG22-47	gb_13	93.3	0.0054	0.05	1.72	0.80	10.4	61.5	644.7	150.3	9919.2	185.0

Table 2.7 continued

Sample	Analysis	Age	La	Pr	Sm	Eu	Gd	Dy	Yb	Lu	Hf	U
IG22-47	gb_14	93.3	0.0068	0.02	0.97	0.44	5.7	41.1	520.1	122.9	9685.5	140.7
IG22-47	gb_15	93.3	0.0180	0.06	2.08	0.94	11.2	64.8	627.8	140.5	10160.2	178.4
IG22-48	gd_43	97.2	0.0666	0.04	1.58	0.31	9.7	72.5	389.9	67.5	9726.8	124.3
IG22-48	gd_44	97.2	0.0428	0.07	3.86	0.81	18.5	128.1	742.8	134.2	9543.4	405.8
IG22-48	gd_45	97.2	0.0037	0.03	2.12	0.52	12.1	91.0	504.7	90.0	9595.3	177.5
IG22-48	gd_46	97.2	0.1910	0.11	3.12	0.76	14.7	98.2	550.3	97.7	8679.0	307.3
IG22-48	gd_47	97.2	0.0388	0.07	3.20	0.50	16.0	108.4	601.7	100.6	10304.1	309.8
IG22-48	gd_48	97.2	0.0318	0.10	5.73	1.10	28.5	181.1	906.7	142.7	10760.8	327.1
IG22-48	gd_49	97.2	0.0071	0.03	2.66	0.53	14.9	97.5	470.7	78.9	9657.3	123.5
IG22-49	gd_29	87.8	0.1320	0.07	2.34	0.33	10.1	73.6	362.0	60.7	10319.6	184.9
IG22-49	gd_30	87.8	0.0268	0.07	2.34	0.44	13.6	97.3	456.4	76.3	10793.2	227.6
IG22-49	gd_31	87.8	0.0090	0.05	3.39	0.50	15.5	107.4	586.0	96.8	10907.7	264.3
IG22-49	gd_32	87.8	9.2000	2.92	6.14	0.74	18.4	112.1	461.0	76.8	10315.3	300.9
IG22-49	gd_33	87.8	0.3380	0.19	2.73	0.52	13.8	102.8	531.2	90.6	10300.9	321.6
IG22-49	gd_34	87.8	0.0128	0.04	3.07	0.48	14.4	105.5	563.9	90.1	10412.4	254.7
IG22-49	gd_35	87.8	0.0980	0.09	3.84	0.58	17.1	120.3	485.7	82.8	9031.6	250.2
IG22-50	gc_8	97.6	0.0276	0.04	2.69	0.46	10.4	111.1	671.9	116.4	10040.0	408.4
IG22-50	gc_9	97.6	0.0197	0.22	14.86	1.64	52.6	429.0	1722.6	262.3	7973.3	1572.6
IG22-50	gc_10	97.6	0.1258	0.49	24.31	2.98	86.9	635.2	1982.9	297.9	9230.8	1305.9
IG22-50	gc_12	97.6	0.0143	0.17	7.60	1.43	26.2	188.7	662.8	102.7	9192.5	364.2
IG22-50	gc_13	97.6	0.0475	0.64	25.84	2.77	87.5	670.8	2636.6	397.7	9946.7	1686.4
IG22-50	gc_14	97.6	0.0529	0.66	28.43	3.39	103.3	758.8	2830.0	427.9	8667.5	1681.0
IG22-51A	jd_81	99.1	0.0041	0.10	7.47	0.84	55.4	286.2	1152.0	194.7	11738.2	368.1
IG22-51A	jd_82	99.1	0.0041	0.12	8.00	1.01	65.1	326.8	1243.4	212.3	11928.0	534.1
IG22-51A	jd_83	99.1	0.0016	0.03	1.85	0.59	11.6	53.1	302.4	58.1	10181.5	71.9
IG22-51A	jd_84	99.1	0.0088	0.14	7.03	0.87	45.7	234.2	1192.2	215.9	11915.5	508.1
IG22-51A	jd_85	99.1	0.0138	0.23	14.22	1.96	106.1	509.6	1844.4	293.0	10416.6	425.2
IG22-51B	gf_64	100.8	0.0303	0.06	3.74	0.45	26.2	156.1	826.0	134.7	11417.1	337.1
IG22-51B	gf_65	100.8	0.3900	0.16	2.97	0.46	21.8	131.3	606.1	107.6	11469.1	373.5
IG22-51B	gf_66	100.8	4.2500	2.53	14.68	4.20	51.0	184.5	671.5	117.8	13659.0	646.2
IG22-51B	gf_67	100.8	0.0090	0.08	5.28	0.79	35.9	209.7	831.3	137.2	10480.3	323.5
IG22-51B	gf_68	100.8	0.0471	0.06	3.19	0.39	23.0	138.7	597.0	103.6	11108.7	294.1
IG22-51B	gf_69	100.8	0.0128	0.02	2.10	0.33	16.1	99.2	521.7	96.6	12389.5	593.2
IG22-51B	gf_70	100.8	0.0253	0.08	4.61	0.76	31.3	187.3	761.1	131.5	9695.4	285.6
IG22-53	gf_57	235.5	0.0683	0.27	6.87	0.33	38.9	207.6	697.8	111.7	10307.8	385.6
IG22-53	gf_59	235.5	0.0940	0.68	23.73	0.71	105.4	438.7	910.0	128.0	9480.3	752.6
IG22-53	gf_60	235.5	0.0038	0.03	4.08	0.16	35.2	236.9	698.1	108.7	12262.3	573.4
IG22-53	gf_61	235.5	0.0490	0.35	14.70	0.60	75.7	321.5	921.6	141.1	10183.4	197.7
IG22-53	gf_62	235.5	0.0015	0.02	3.08	0.11	26.4	249.1	1342.6	210.3	12980.6	883.1
SAN01	jc_6	90.0	0.0005	0.09	5.34	1.26	27.3	144.5	593.0	107.8	9806.1	190.8
SAN01	jc_7	90.0	0.0073	0.11	4.63	1.27	28.7	194.3	1090.3	210.3	9659.9	773.7
SAN01	jc_8	90.0	0.1620	0.07	3.19	1.10	16.7	91.4	414.0	75.2	9692.4	167.4
SAN01	jc_9	90.0	0.0429	0.04	2.14	0.57	9.9	60.2	340.6	68.6	9229.5	157.4
SAN01	jc_10	90.0	0.3470	0.56	5.81	1.84	23.5	139.1	613.9	116.1	10545.1	1164.2
SAN02	jc_31	45.2	0.0028	0.02	0.92	0.41	5.4	29.0	188.5	40.1	9059.7	100.6
SAN02	jc_33	45.2	0.8430	0.37	6.10	0.51	32.4	167.3	678.1	113.8	9868.4	406.6

Table 2.7 continued

Sample	Analysis	Age	La	Pr	Sm	Eu	Gd	Dy	Yb	Lu	Hf	U
SAN02	jc_34	45.2	0.0027	0.03	1.64	0.19	13.4	83.3	380.7	70.9	9572.7	374.0
SAN02	jc_35	45.2	0.0460	0.06	3.33	0.32	21.5	128.6	575.5	100.8	9219.5	447.3
SAN03	jc_36	44.1	0.0042	0.03	2.35	0.16	15.9	103.7	508.1	92.1	10542.2	398.9
SAN03	jc_37	44.1	4.4600	1.58	8.45	0.28	44.9	235.2	988.9	167.5	10480.5	908.0
SAN03	jc_38	44.1	0.0075	0.04	2.92	0.17	21.6	132.2	641.3	115.7	10924.6	540.2
SAN03	jc_39	44.1	0.0604	0.05	1.83	0.09	12.9	80.5	363.8	67.3	8009.9	429.0
SAN03	jc_40	44.1	0.6410	0.27	2.30	0.36	14.9	92.9	473.5	87.9	9804.8	355.1
SAN04	jc_26	46.7	0.0038	0.01	0.93	0.10	7.6	52.2	312.2	57.9	8718.4	285.8
SAN04	jc_27	46.7	0.0455	0.05	3.26	0.24	22.5	139.6	624.4	108.2	11011.2	476.9
SAN04	jc_28	46.7	0.4150	0.16	2.58	0.15	17.2	109.9	513.8	93.4	9873.6	482.5
SAN04	jc_30	46.7	1.4120	0.54	4.75	0.25	29.8	168.3	661.0	118.3	9282.8	586.8
SAN05	jc_21	42.0	0.0237	0.21	8.89	1.36	48.0	236.1	806.6	136.6	9647.7	333.6
SAN05	jc_22	42.0	0.2600	0.25	5.17	0.89	27.2	146.8	583.1	104.9	9331.8	269.0
SAN05	jc_23	42.0	0.0341	0.37	14.21	2.30	83.0	402.3	1263.1	219.6	8980.2	642.5
SAN05	jc_24	42.0	0.0207	0.19	7.92	1.23	38.3	185.4	601.0	103.2	10291.4	170.2
SAN05	jc_25	42.0	0.0338	0.08	3.75	0.77	23.4	124.7	482.4	88.4	9479.1	157.3
SAN06	jc_1	43.6	0.1014	0.06	2.53	0.39	13.9	82.4	359.2	58.0	10512.0	103.6
SAN06	jc_2	43.6	0.0006	0.03	2.41	0.43	16.4	93.6	340.4	57.9	10931.8	109.1
SAN06	jc_3	43.6	0.0244	0.25	8.42	1.48	43.3	232.3	711.4	120.8	9233.9	203.3
SAN06	jc_4	43.6	0.0001	0.03	2.88	0.52	16.7	95.3	342.0	58.4	10220.1	92.9
SAN06	jc_5	43.6	0.0019	0.08	3.96	0.65	20.8	121.2	491.6	86.7	9276.2	200.5
SAN08	jc_41	20.2	0.0035	0.10	4.97	0.75	24.2	109.9	340.1	56.0	10568.0	118.4
SAN08	jc_42	20.2	0.0107	0.08	3.21	0.56	17.4	87.1	301.1	52.0	11366.4	114.1
SAN08	jc_43	20.2	0.0003	0.06	3.06	0.52	17.8	86.4	311.5	54.4	10970.9	129.6
SAN08	jc_44	20.2	0.0171	0.21	9.44	1.51	44.0	190.4	567.6	96.9	9995.9	212.4
SAN08	jc_45	20.2	0.0022	0.02	2.05	0.39	13.5	69.8	254.3	44.3	11927.2	104.8
SAN09	jc_46	20.2	0.0095	0.13	5.64	0.99	30.2	129.9	403.3	66.1	10383.8	143.4
SAN09	jc_47	20.2	0.0076	0.03	1.75	0.36	11.1	59.0	216.8	36.6	11208.5	88.6
SAN09	jc_48	20.2	0.0035	0.10	5.67	0.91	28.0	119.3	362.8	61.5	10060.6	121.5
SAN09	jc_49	20.2	0.1040	0.16	6.34	1.06	33.7	163.0	510.7	86.8	10619.7	181.4
SAN09	jc_50	20.2	0.0027	0.10	6.08	1.18	31.1	140.4	466.2	81.3	11732.3	135.7
SAN10	jc_51	29.1	0.0439	0.16	5.43	0.88	28.1	123.3	363.7	60.7	9404.1	73.0
SAN10	jc_52	29.1	3.8300	1.53	6.16	0.92	26.2	120.0	430.0	74.9	10467.3	118.5
SAN10	jc_53	29.1	0.0172	0.28	9.96	2.35	56.7	247.2	729.9	118.2	8858.6	133.2
SAN10	jc_55	29.1	0.0077	0.10	5.31	0.99	28.8	122.8	400.5	66.8	9991.0	74.5
SAN13	jc_61	34.3	0.4000	0.29	6.59	0.19	53.4	369.4	1686.4	286.4	14780.1	1357.8
SAN13	jc_62	34.3	2.1600	1.04	5.92	0.74	31.7	152.5	603.9	104.8	11438.4	315.8
SAN13	jc_63	34.3	0.1056	0.09	3.54	0.38	27.3	166.0	665.2	113.5	12037.5	445.3
SAN13	jc_64	34.3	0.0008	0.06	2.99	0.62	17.7	82.6	347.2	61.2	9406.7	90.0
SAN14	jc_11	33.7	0.0059	0.04	3.37	0.20	28.4	220.8	1089.6	189.7	13855.5	1020.7
SAN14	jc_12	33.7	4.5300	2.26	15.62	0.54	64.0	374.1	1477.3	266.7	12345.5	1297.3
SAN14	jc_13	33.7	6.1700	1.86	5.73	0.36	30.9	215.7	1002.2	181.5	12301.6	743.0
SAN14	jc_14	33.7	0.0045	0.01	1.93	0.13	16.3	141.9	754.1	138.1	14003.4	560.0
SAN14	jc_15	33.7	0.0319	0.06	4.80	0.20	39.1	306.3	1332.0	239.8	13596.4	1338.2
SAN15	jc_17	33.7	0.0910	0.09	3.49	0.64	20.3	122.7	570.6	104.9	10410.0	233.2
SAN15	jc_18	33.7	0.2000	0.12	3.84	0.23	31.0	222.1	880.0	160.8	10809.1	693.8

Table 2.7 Continued

Sample	Analysis	Age	La	Pr	Sm	Eu	Gd	Dy	Yb	Lu	Hf	U
SAN15	jc_19	33.7	0.0531	0.08	4.87	0.42	35.0	242.3	995.0	180.0	10968.0	804.2
SAN15	jc_20	33.7	0.3450	0.31	5.28	0.34	38.8	294.2	1351.3	226.1	13918.6	1212.2
SAN16	jc_66	33.6	0.1330	0.07	2.07	0.17	22.6	183.0	1046.7	185.1	12041.8	1037.4
SAN16	jc_67	33.6	0.0951	0.14	4.77	0.29	40.4	303.2	1405.7	245.5	12596.4	1116.3
SAN16	jc_68	33.6	0.0038	0.03	3.52	0.15	31.2	217.7	1010.5	179.4	13852.2	918.6
SAN16	jc_69	33.6	0.0025	0.01	1.20	0.11	12.2	105.4	661.3	117.7	13563.7	689.6
SAN16	jc_70	33.6	0.0169	0.14	6.39	1.53	38.2	172.4	622.4	110.3	9981.9	196.3
SAN17	jc_71	40.2	0.0013	0.02	0.99	0.35	7.7	41.8	229.6	43.0	10366.5	102.2
SAN17	jc_72	40.2	0.0031	0.04	2.17	0.54	14.0	69.4	346.5	64.5	10662.8	167.3
SAN17	jc_73	40.2	0.0042	0.04	1.91	0.67	11.8	59.8	336.2	61.1	10204.5	187.3
SAN17	jc_74	40.2	0.0116	0.08	2.78	0.59	11.3	41.1	225.0	44.5	11122.4	191.2
SAN17	jc_75	40.2	0.0361	0.08	3.13	0.87	15.6	68.5	314.4	58.8	9662.6	116.7
SAN18	jc_76	139.8	0.0671	0.50	9.92	0.90	37.8	127.3	444.0	79.4	10372.1	802.4
SAN18	jc_77	139.8	0.0426	0.48	11.05	0.88	40.9	138.0	489.7	87.4	10823.3	833.3
SAN18	jc_78	139.8	0.0023	0.03	1.14	0.15	5.8	24.8	107.3	18.6	9471.9	95.1
SAN18	jc_79	139.8	0.0171	0.24	6.11	0.71	24.6	78.8	242.2	40.2	9296.1	175.9
SAN18	jc_80	139.8	0.0255	0.42	8.94	0.99	33.5	102.8	290.1	47.7	8931.1	262.0
SAN19	jd_1	19.9	0.0192	0.18	7.17	1.21	28.7	127.6	333.6	55.6	9964.4	103.9
SAN19	jd_2	19.9	0.0021	0.03	2.70	0.52	17.4	103.3	355.4	59.2	11235.2	115.1
SAN19	jd_3	19.9	0.0036	0.05	4.22	0.85	24.7	125.3	370.1	58.3	10815.7	114.2
SAN19	jd_4	19.9	0.0040	0.02	1.92	0.32	9.6	61.1	240.1	42.1	11106.0	86.1
SAN19	jd_5	19.9	0.0005	0.03	1.54	0.24	9.3	51.9	205.0	35.4	11297.7	86.1
CM11	P1143	462.5	0.0131	0.04	3.18	1.13	21.8	98.3	387.1	84.5	9534.0	287.8
CM11	P1136	474.4	0.0121	0.11	3.52	1.52	18.5	61.2	184.2	41.3	8214.8	108.9
CM11	P1146	682.5	0.0144	0.05	2.30	0.93	11.0	44.7	201.8	47.8	8408.2	76.7
CM16	P1626	234.8	0.0052	0.06	3.23	0.24	21.1	115.8	477.2	99.6	7324.8	141.7
CM16	P165	255.9	0.0149	0.03	1.74	0.18	12.3	73.5	426.6	95.4	10371.0	316.8
CM16	P1641	413.7	0.2244	0.22	1.74	0.39	9.8	58.7	300.0	63.5	13428.9	652.5
CM16	P1633	566.8	0.0193	0.46	9.87	0.28	25.4	32.9	17.8	2.8	12364.5	285.9
CM16	P1649	572.1	0.0025	0.06	1.89	0.69	7.5	22.9	81.0	16.4	11432.9	235.1
CM16	P167	644.6	0.7420	0.52	6.53	1.36	26.0	77.7	195.7	38.7	7069.1	298.3
CM16	P1620	850.6	0.0043	0.05	1.85	0.25	11.4	48.9	176.3	36.5	7394.1	88.5
CM16	P1638	923.6	-0.0001	0.02	1.95	0.11	12.3	56.9	198.5	39.1	11030.2	241.2
CM16	P1634	1126.8	0.0106	0.08	4.09	0.58	29.2	172.9	725.6	147.5	11885.6	270.2
CM3	P345	604.1	0.0292	0.04	1.48	0.19	9.1	39.0	123.2	25.3	8514.4	89.8
CM3	P336	619.6	0.0480	0.08	4.19	0.16	26.3	112.3	327.4	63.0	10732.9	321.6
CM5	P510	229.2	0.0440	0.08	4.80	0.12	41.5	267.1	974.4	187.0	10733.5	344.0
CM5	P530	572.5	0.0094	0.03	2.33	0.30	16.9	108.7	531.8	111.9	11143.8	184.8
CM7	P742	393.9	-0.0001	0.01	0.79	0.33	5.4	22.0	105.5	25.1	12052.4	117.1
CM7	P737	519.0	0.0026	0.05	3.11	0.13	20.0	70.5	79.4	13.2	8808.2	60.1
CM7	P736	665.6	0.0000	0.02	1.17	0.49	7.2	35.7	182.8	42.8	7811.9	47.8
CM7	P730	1033.8	0.0249	0.06	1.95	0.77	12.4	57.7	282.2	64.2	10060.2	211.4
CM7	P746	220.1	0.1251	0.16	6.01	2.09	29.7	89.6	291.4	64.5	12125.7	535.2
CT004	ge_27	472.4	0.0151	0.08	2.55	0.78	10.7	63.3	372.7	66.5	9933.1	135.8
CT004	ge_28	450.6	0.0809	0.35	9.69	0.31	42.3	269.6	921.2	138.6	9919.2	235.8
CT004	ge_29	812.2	0.0675	0.15	5.86	0.34	33.6	225.5	658.2	94.3	11195.8	407.2

2.7 AFT methodology

Fission tracks are microscopic zones of damage created as fission fragments traverse a solid material. While several naturally occurring isotopes undergo spontaneous nuclear fission, only ^{238}U generates a significant number of tracks over geological time scales (Naeser, 1976). Fission track (FT) dating utilizes these damage features to record geological events. This technique includes two main approaches: multi-grain methods and single-grain methods, with the latter divided into the external detector method (EDM) and LA-ICP-MS method (Hurford & Green, 1983).

In EDM, the ratio of spontaneous to induced tracks is combined with the spontaneous fission decay rate of uranium and a physical constant to calculate the time over which fission tracks have accumulated in a mineral. Under certain conditions, this time corresponds to the sample's age (Hurford & Green, 1983). The method involves measuring ^{238}U spontaneous fission tracks in the grain itself and counting ^{235}U induced tracks in a detector with low uranium content, such as muscovite mica. During irradiation, the detector is positioned against the grain mount, enabling a precise comparison between the grain and its "mirror image" in the detector (Hurford & Green, 1983).

The age determination in EDM uses Equation 2.5 (additional information can be found in Vermeesch, 2019) involving several parameters: λ_D , the total decay constant of ^{238}U ($1.55125 \times 10^{-10} \text{ yr}^{-1}$; Jaffey et al., 1971); ζ , a calibration factor (Hurford & Green, 1983); ρ_s , the density of spontaneous fission tracks on the crystal surface; ρ_i , the density of induced fission tracks in the mica detector; and ρ_d , the density of induced fission tracks in a dosimeter glass with known uranium concentration.

$$t = \frac{1}{\lambda_D} \ln \left(1 + \frac{1}{2} \lambda_D \zeta \rho_d \frac{\rho_s}{\rho_i} \right) \quad \text{equation 2.5}$$

Apatite fission track (AFT) preparation for the external detector method (EDM) was carried out at the Thermochronology Laboratory of the University of Padova, where 32 aliquots were prepared for analysis. Apatite crystals were selected using a binocular Olympus SZX16 microscope, mounted on double-sided tape, embedded in epoxy sheets, and polished to expose their internal surfaces. The exposed surfaces were etched with nitric acid HNO_3 for 20 seconds at ambient temperature to reveal the fission tracks, as HF preferentially dissolves the damaged areas created by the spontaneous fission of uranium atoms within the crystal lattice (Hurford & Green, 1983). The apatite mounts were then covered with mica detectors and irradiated during two separate campaigns (OSU28 and OSU29B) at the Oregon State Radiation Center, USA, at nominal fluences of $1.40 \times 10^6 \text{ n/cm}^2$ and $1.22 \times 10^6 \text{ n/cm}^2$, respectively. The irradiations also included Fish Canyon, Mount Dromedary, and Durango age

standards, as well as CN5 dosimeter glasses. Following irradiation, the mica detectors were detached, etched with 40% HF at 20°C for 45 minutes, and prepared for track counting.

Fission tracks were counted dry at 1250× magnification using an Olympus BX51 optical microscope with an automated DeltaPix system at the University of Padova. Ages were calculated using a zeta value of 356.04 ± 14.58 . AFT ages are reported as central ages, calculated using the radial plot software developed by Galbraith (1990), and implemented by Vermeesch (2019) in Isoplot R. Detailed results and analytical parameters are provided in Table 2.8.

Table 2.8 AFT data

Sample	nG	pd (10^5 /cm ²)	Ns	Ni	P(χ^2) (%)	Central Age (Ma)	$\pm 1\sigma$ (Ma)	U (ppm)	nL	MTL	$\pm 1\sigma$ (μ m)	Dpar (μ m)
IG22-02	20	12.06	75	1004	100	15.40	1.90	0.18	29	14.18	1.18	2.90
IG22-05	8	11.99	21	150	100	29.80	7.10	0.08	24	13.43	1.51	2.30
IG22-04	19	11.92	55	495	100	23.50	3.50	0.53	22	13.71	1.64	2.20
IG22-07	20	11.78	166	1278	100	27.20	2.50	0.25	30	13.70	1.40	2.00
IG22-08	20	11.71	521	2689	90	40.30	2.50	0.61	72	13.23	0.90	1.80
IG22-09	20	11.64	368	2988	100	25.50	1.80	0.63	36	13.71	1.43	2.30
IG22-13	20	11.57	33	341	100	19.90	3.70	0.04	30	13.82	1.20	2.02
IG22-14	10	12.40	16	129	100	23.10	6.6	0.03	24	13.06	0.80	3.00
IG22-15	3	12.33	6	35	75	38.00	17.0	0.06	-	-	-	
IG22-17	5	12.25	11	71	90	34.00	11.0	0.08	-	-	-	
IG22-18	19	11.50	110	713	100	31.50	3.50	0.15	14	14.15	1.09	2.80
IG22-19	13	12.18	16	97	100	35.90	9.7	0.05	31	13.37	0.92	2.20
IG22-21	20	12.10	331	1966	91	36.20	2.60	0.57	71	12.10	1.60	3.10
IG22-23	20	12.03	178	1214	100	31.30	2.80	0.35	3	13.89	0.90	2.80
IG22-26	11	11.23	66	456	100	28.90	4.00	1.62	21	12.94	1.43	1.90
IG22-30	20	11.95	348	1944	88	38.00	2.70	0.43	73	12.93	1.20	2.34
IG22-34	19	11.81	86	781	100	23.10	2.80	0.21	23	13.80	1.30	2.54
IG22-35	19	11.09	46	297	100	30.50	5.00	0.07	-	-	-	2.11
IG22-37	16	10.95	63	290	100	42.50	6.10	0.08	21	12.93	1.50	2.32
IG22-38	20	10.88	36	216	100	32.10	5.90	0.03	46	13.57	1.30	2.74
IG22-39	18	11.74	41	373	100	22.70	2.90	0.12	23	13.23	1.70	3.01
SAN01	20	10.74	80	450	100	33.80	4.30	0.07	39	13.06	1.10	2.41
IG22-43	20	10.67	429	2594	84	31.30	2.10	0.61	37	12.79	1.10	2.22
IG22-44	20	11.58	715	2226	97	65.80	4.00	0.29	48	12.49	1.30	2.05
IG22-45	19	10.61	157	719	99	41.00	4.00	0.19	28	13.26	1.20	2.49
IG22-46	14	10.54	218	1165	99	35.00	3.00	0.36	20	12.30	1.00	1.85
IG22-47	19	10.47	578	2193	96	48.80	3.10	0.58	48	13.36	1.00	1.75
IG22-49	20	10.40	772	3043	88	46.70	2.80	0.72	52	12.91	1.40	1.82
IG22-50	11	10.33	98	443	100	40.30	4.80	0.20	7	12.01	0.90	2.29
IG22-51-A	18	10.26	81	346	100	42.60	5.50	0.10	29	12.78	0.90	2.04
IG22-51-B	20	10.19	187	1283	98	26.30	2.40	0.32	37	12.89	1.40	2.12
IG22-53	20	10.12	432	2185	83	35.50	2.40	0.47	33	13.30	1.40	1.81

2.8 3D Thermokinematic-modelling (Pecube)

To reconstruct the thermal and exhumation history of two regions in southern Ecuador—(i) the Southern Real Cordillera (SRC), comprising the Sabanilla Unit and the Zamora Batholith, and (ii) the La Bocana and La Victoria units of the Amotape-Tahuin Block (ATB)—we employed the three-dimensional thermo-kinematic modeling code Pecube (Braun et al., 2012). The exhumation history was modeled as a series of discrete phases, each defined by a constant exhumation rate between specified start and end times. To optimize computational efficiency, the topographic surface was simplified by resampling the Shuttle Radar Topography Mission (SRTM) digital elevation model to a resolution of 1 km—sufficient for accurate modeling results, as demonstrated by Valla et al. (2011).

The model setup for the SRC zone comprised a rectangular domain measuring 29×37 km, with a vertical fault (90° dip) of 15 km depth positioned outside the modeled block, to model the exhumation of rocks. The boundary conditions were a basal temperature ranging from 300 to 900 °C, a surface temperature of 24 °C and an atmospheric lapse rate of 6 °C/km. The heat production rate was permitted to vary between 0 and 100 $\mu\text{W m}^{-3}$, while the model thickness was fixed at 55 km.

During our modeling processes, positive rates (+) are interpreted as periods of exhumation and consequent cooling. While, negative rates (–) signify more complex thermal scenarios where samples experienced apparent reheating, thermal overprinting or isotherm relaxations. This last can be potentially driven by magmatic heating, lithospheric thinning, or burial episodes.

The inversions were run with one to four phases of exhumation ($Pe = 1$ to 4). The first pulse was constrained to be positive, representing a phase of forced exhumation. The subsequent three pulses were modeled with variable rates ranging from -2 to 2 km/My (either positive related to exhumation, or negative associated with thermal relaxation). The model is additive, meaning that exhumation rates from the second pulse onward build upon those of the preceding phases. In this framework, positive rates (+) are interpreted as periods of exhumation and consequent cooling. While, negative rates (–) signify more complex thermal scenarios where samples experienced apparent reheating. This typically results from the relaxation of isotherms towards steady-state following a thermal perturbation, potentially driven by magmatic heating, lithospheric thinning, or burial episodes.

For the ATB zone, the model comprises a rectangular area measuring 37×23 km, incorporating a thrust fault at the center of the block, extending down to a maximum depth of 15 km, this fault separates the La Bocana unit (Block 1) from the La Victoria unit (Block 2). The fault is represented with a 30° dip, consistent with a moderate-angle thrust geometry. Although the possibility of a pre-existing normal fault could imply a steeper dip, 30° was selected as it corresponds to the mechanically optimal

angle for thrusting, following Andersonian fault theory (Anderson, 1905). However, varying the fault angle did not result in significant differences in the modeled exhumation rates. The thermal boundary conditions were identical to those of the SRC zone: a basal temperature range of 300–900 °C, a surface temperature of 24 °C and a lapse rate of 6 °C/km. The heat production range remained at 0–100 $\mu\text{W m}^{-3}$, while the model thickness was fixed at 45 km. Table 2.9 summarizes the fixed thermal and flexural parameters used in Pecube, while Table 2.10 lists the variable parameters.

Table 2.9 Fixed imposed boundary conditions and material properties used for thermo-kinematic models (Pecube)

Parameter	Symbol	Value	Unit
Crustal density ¹	ρ_c	2700	kg m^{-3}
Mantle density ¹	ρ_m	3200	kg m^{-3}
Young's modulus	E	70	Gpa
Poisson ratio ¹	ν	0.25	Dimensionless
Model thickness ²	L	45 (for ATB) 55 (for SRC)	Km
Effective elastic thickness ³	T_e	30	Km
Thermal diffusivity ¹	$\kappa = k / (\rho m^c)$	25	$\text{Km}^2 \text{my}^{-1}$
Thermal conductivity ¹	k	25	$\text{Wm}^{-1}\text{K}^{-1}$
Surface temperature	T_0	24	°C
Atmospheric lapse rate ¹	$\beta = -dT/dz$	6	°C Km ⁻¹
Starting time for exhumation ^{4,5}	t_s	190	Ma

Subscripts indicate reference for some parameters: 1: Braun et al. (2012). 2: Araujo (2016). 3: Pérez-Gussinyé et al. (2007). 4: Spikings et al. (2005). 5: Spikings et al. (2010).

Table 2.10 Parameters variable for PeCube inverse modeling

Parameter	Symbol	Value	Unit
Temperature at the base of the model	H	300 to 900	μWm^{-3}
Heat production	T_s	0 to 100 250 to 0 (for the ATB)	$\mu\text{W m}^{-3}$
Time for exhumation	t_{Ei}	200 to 0 (for the SRC)	Ma
Number of phases of exhumation	i	1 to 4	
Exhumation rate for each phase i	E_i	-2 to 2	Km Ma^{-1}

2.9 2D Thermal and exhumation history of intrusions (4DTherm)

To quantify the exhumation histories of those samples which present concordant or nearly identical AFT and U–Pb zircon ages we employed the 4DTHERM code (Fu et al., 2010). 4DTHERM solves the governing equations of heat and mass transfer in a magmatic body using an explicit finite difference scheme. The model accounts for conductive and convective cooling, as well as erosional exhumation. Models were run until achieving consistent results, with successful convergence in general matching mode. Table 2.11 summarizes both the fixed boundary conditions and the variable parameters employed in the software.

Table 2.11 List of paraments, symbols, and values used in the 4Dtherm modelling

Parameter	Symbol	Value	Unit
Acceleration of gravity	g	9.8	m/s ²
Basal heat flow	Q _{basal}	65	mW/m ²
Density of water	ρ _f	1000	kg/m ³
Density of magma	ρ _m	2700	kg/m ³
Density of rock (sedimentary)	ρ _r	2300	kg/m ³
Dynamic viscosity of magma	μ _m	10 ¹²	Pas
Dynamic viscosity of water	μ _f	2.82 X10 ⁻⁴	Pas
Emplacement depth of magma	D _{emplace}	Variable	m
Geothermal gradient	dT/dy	Variable	°C/Km
Erosion rate of country rock	v _c	Variable	mm/yr
Erosion rate of igneous rock	v _i	Variable	mm/yr
Initial magma temperature	T _m	1000	°C
Latent heat of magma	L _m	4.18 x 10 ⁵	J/Kg
Permeability of rock	κ _r	1.0x10 ⁻¹⁵	m ²
Porosity of rock	Φ	1%	
Specific heat of magma	C _m	Variable	J/Kg°C
Specific heat of rock	C _r	Variable	J/Kg°C
Specific heat of water	C _f	4180	J/Kg°C
Surface temperature	T _s	15-25	°C
Thermal conductivity of rock	λ _r	2.5	W/m°C
Thermal conductivity of magma	λ _m	3	W/m°C
Volumetric coefficient of thermal expansion of water	α _f	10 ⁻³	1/°C
Volumetric coefficient of thermal expansion of magma	α _m	2.0x10 ⁻⁵	1/°C

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CHAPTER 3: PERVASIVE CRUSTAL REWORKING ALONG THE PROTO-ANDEAN MARGIN OF NORTHERN PERU AND SOUTHERN ECUADOR: INSIGHTS FROM U-PB, GEOCHEMICAL, AND ISOTOPIC ANALYSES OF ZIRCON

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Declaration statement

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CRedit authorship contribution statement

Sandoval Espinel, L.C: Conceptualization, Investigation, Data curation, Methodology, Writing – original draft. **Witt, C:** Conceptualization, Supervision, Writing – review & editing. **Zattin, M:** Conceptualization, Supervision, Writing – review & editing. **Poujol, M:** Methodology, Data curation, Writing – review. **Bruguier, O:** Methodology, Data curation, Writing – review. **Chiaradia, M:** Writing – review & editing. **Bermudez, M:** Writing – review & editing.

HIGHLIGHTS

1. Zircon grains (1000–200 Ma) formed via repeated crustal reworking episodes.
2. Increased juvenile input (380–310 Ma), followed by sustained high $\delta^{18}\text{O}$ values until 220 Ma.
3. Evolved $\varepsilon_{\text{Hf}(t)}$ (-10 to -30) at 640–510 Ma suggests ancient Amazonian craton input.
4. U-Pb ages match central Andes spectra but show stronger northern Gondwana affinities.

ABSTRACT

Continental growth at convergent margins is the result of complex interactions between mantle additions and crustal reworking. This study examines the pre-Andean tectono-magmatic evolution of the central-northern Andean margin ($\sim 1^{\circ}\text{S}$ to $\sim 6^{\circ}\text{S}$) using U-Pb geochronology, trace element analysis, and $\epsilon\text{Hf}_{(t)}$ and $\delta^{18}\text{O}$ isotopic data from zircons embedded within detrital and bedrock samples. U-Pb dating revealed two primary clusters: a Late Neoproterozoic population (640–550 Ma) associated with the Brasiliano/Pan-African orogeny and a Permian–Triassic population (280–200 Ma) associated with the Gondwanides orogeny. There were also two minor clusters at ~ 960 Ma and ~ 345 Ma. Zircon grains dating between 1000–200 Ma crystallized during repeated episodes of crustal reworking, characterized by highly evolved $\epsilon\text{Hf}_{(t)}$ signatures (-10 to -30) at 640–510 Ma that are indicative of ancient Amazonian cratonic sources, with no evidence of Pampean or Famatinian contributions. The results also show a marked increase in juvenile input during the Carboniferous (380–310 Ma), followed by predominantly continental input as indicated by persistently high $\delta^{18}\text{O}$ values up to 220 Ma. Although the U-Pb age spectra of these samples are similar to that of the central Andes, isotopic and geochemical data reveal that the central-northern Andean margin evolved within a tectonic context that is more closely aligned with central-eastern Gondwana (i.e., central Africa and eastern South America) rather than typical Andean settings.

Keywords: Crustal evolution, Hf isotopes, S-type granitoids, U-Pb LA-ICP-MS dating, zircon geochemistry, Brasiliano Orogeny.

3.1 Introduction

The growth of continents is a first-order issue in the Earth sciences. Along convergent margins, continental growth is mostly attributed to the transfer of material from the subducting plate to the overriding plate, whereas crustal growth (i.e., the formation of new crust) is typically attributed to magmatism, through the addition of mantle melts and subsequent fractional crystallization, magma mixing and assimilation processes (e.g., Couzinié et al., 2016). However, magmatism can also be associated with melting processes of the upper plate, which act as a mechanism by which the pre-existing crust is further refined (e.g., Hagen-Peter & Cottle, 2018). Determining the proportion of new additions from mantle-derived magmas compared to reworked material is, therefore, a central aspect of geodynamics and regional tectonic models (Belousova et al., 2010; Kemp et al., 2006). At present, the primary way of approaching this issue is by the analysis of zircon petrochronology (e.g., U-Pb geochronology, trace elements, and isotopic data) in subduction-related tectonic settings, where

continental growth mechanisms are well-known and magmas represent the average composition of the continental crust (Hawkesworth et al., 2013).

The Central-North Andean margin (CNAM), a region that spans central–southern Ecuador and northern Peru, is an exceptional natural laboratory for studying crustal evolution and their associated magmatic processes because it encompasses a wide temporal range of Andean history from the Paleoproterozoic-early Paleozoic to the recent Cenozoic era (Chew et al., 2008; Mišković et al., 2009; Pepper et al., 2016). Furthermore, the location of the CNAM is crucial in the context of the evolution of Gondwana as it is situated at the junction between the southern proto-Andean margin (Terra Australis)—which was influenced by both the Pampean and Famatinian orogenies—and the northern proto-Andean margin, which was influenced by the Brasiliano/Pan African belt and the opening of the Rheic Ocean (Nance et al., 2012; Oriolo et al., 2021; Ramos, 2018; Schwartz et al., 2008; Van der Lelij, 2013). Nevertheless, despite its long and complex history, research in this area has primarily focused on understanding the Andean cycle (i.e., the geological history of the last 200 Ma), leaving much of earlier pre-Andean subduction history comparatively underexplored. In particular, the interval between 620–460 Ma represents one of the most significant magmatic and tectonic phases along the proto-Andean margin; specifically, it marks the onset of a prolonged subduction process along the paleo-Pacific margin of Gondwana (Capaldi et al., 2021; Chew et al., 2007). This period includes the Brasiliano/Pan-African orogeny, a process that involved the convergence and amalgamation of multiple cratons, ultimately leading to the formation of the Gondwana supercontinent (Oriolo et al., 2021; Santos et al., 2008). It also includes the Pampean orogeny, which was initiated by the eastward subduction of oceanic crust around 555 Ma and culminated in a ridge-trench collision at ca. 525 Ma (Mišković et al., 2009; Schwartz et al., 2008).

Zircon is a highly resistant mineral that can withstand erosion, transport, and the high temperature and pressure conditions associated with metamorphism and melting. It is the most commonly used mineral in provenance models due to its relative ease of dating via the U-Pb isotopic system, its ability to record both mantle and crustal isotopic signatures (mainly via the Lu-Hf and O isotopic systems), and its ubiquity within crustal rocks and mantle xenoliths, among other aspects (Belousova et al., 2002; Grimes et al., 2015). Indeed, detrital zircons may contain a more complete record of magmatic events than preserved arc exposures. In addition to preserving a record of the provenance of clastic sediments, zircon analyses have become a central component of investigations into plate tectonic processes and basin evolution in central and north Andean settings (Jara et al., 2021; Mišković et al., 2009; Spikings et al., 2015).

This study aims to provide insights into the Neoproterozoic to Triassic (~1000–200 Ma) magmatic history and associated crustal evolution of the CNAM by analyzing the petrochronologic record of zircons hosted in Paleozoic and Triassic metamorphic and granitic rocks as well as in late Cretaceous–Cenozoic forearc sedimentary rocks. We analyze the major clusters observed in the U-Pb age spectra as well as zircon geochemistry (such as tectonic discrimination and magmatic reservoir evolution) and relate them to known geodynamic episodes that occurred within the proto-Andean margin. This study also underscores the unique value of Hf and O isotopic signatures in distinguishing overlapping U-Pb age melting sources in zircon analyses, thereby providing invaluable regional insights into both source-to-sink and crustal evolution processes, which could be used to address contemporary geodynamic models, especially those focused on mantle-crust interactions and the evolution of the crust at convergent margins.

3.2 Geological and geodynamic setting

The oldest rock exposures in the CNAM are metamorphic rocks and metasediments that outcrop in both Ecuador and Peru within the Western, Eastern, and Coastal Cordilleras. These exposures comprise different terranes and geological units, including the Olmos terrane, the Olmos-Loja Complex, the Tahuin Block, the Sabanilla Unit, and the Marañon Complex (Fig. 3.1A; Chew et al., 2008; Pepper et al., 2016; Witt et al., 2017). There are only a few constraints on these Paleozoic and metamorphic units in southern Ecuador and northern Peru particularly regarding the tectonic processes they underwent and their roles within the broader geodynamic framework of the region (Aspen et al., 1992; Bahlburg et al., 2020; Chew et al., 2007; Litherland, 1994; Panca et al., 2024; Pratt et al., 2005; Reimann Zumsprekel et al., 2015; Riel et al., 2018; Riel et al., 2014; Spikings et al., 2021; Witt et al., 2017). For instance, Chew et al. (2007) and Spikings et al. (2021) discuss the role of Paleozoic basement units in Andean orogenesis, while Bahlburg et al. (2020) and Panca et al. (2024) integrate provenance analysis within a Paleozoic-Precambrian geodynamic context. While, Witt et al. (2017) examines ancient magmatism and its implications for crustal evolution.

Much of the pre-Andean evolution of the CNAM can either be directly or indirectly ascribed to: i) Neoproterozoic magmatism associated with the Sunsás orogeny that resulted from the collision of Laurentia with Amazonia during the assembly of Rodinia (Cawood & Pisarevsky, 2017; Mišković et al., 2009); ii) a Tonian to Cryogenian rifting episode that marked the initial separation of Laurentia from Amazonia during the breakup of Rodinia, leading to the opening of the Iapetus Ocean (Cawood, 2005; Mišković et al., 2009); iii) Late Neoproterozoic to Early Cambrian magmatism closely associated with the Brasiliano/Pan-African orogeny (600–500 Ma), which marked the convergence and collision of the African and South American continental blocks (Oriolo et al., 2021; Santos et al.,

2008)—simultaneously, the Pampean orogeny (540–520 Ma) represents the initiation of the eastward-dipping subduction of proto-Pacific oceanic lithosphere along the paleo-Pacific margin of western Gondwana (Chew et al., 2007; Mišković et al., 2009); iv) a Cambrian to Ordovician pulse of orogenic and magmatic activity related to the Famatinian orogenic cycle (490–460 Ma), which resulted from the collision of the Paracas Terrane (Chew et al., 2016; Ramos, 2018); v) a passive margin setting that existed during Late Silurian to Early Carboniferous (Chew et al., 2007; Mišković et al., 2009); vi) a Middle Mississippian compressional tectonic regime linked to the closure of the Rheic Ocean resulting from the subduction of Panthalassian oceanic crust (Boekhout et al., 2018; Chew et al., 2007) that culminated in the assembly of the Pangea supercontinent (Boekhout et al., 2018); and vii) the Triassic extensional setting, which has been interpreted as a rift regime that drove seafloor spreading between Central and South America (Carrillo et al., 2021; Chew et al., 2016; Jaillard et al., 1995; Jaillard et al., 1990; Spikings et al., 2015). This phase in western Gondwana was followed by a Late Triassic thermal sag associated with continued rifting (Carrillo et al., 2021). However, Riel et al. (2018) proposed a different interpretation, suggesting that the Permian-Triassic evolution of western Gondwana was primarily influenced by the reorganization of the Tethyan realm. According to their model, the rotation of Pangea led to a decrease in convergence velocity between South America and Panthalassa, which in turn triggered slab steepening, high heat flow in the mantle wedge, and widespread magmatism.

The CNAM exhibits very well-preserved forearc sedimentary basins (Aizprua et al., 2020; Witt & Bourgois, 2010; Witt et al., 2019) that were formed as the result of long-lived subsidence that lasted from the Cretaceous to the Quaternary (Jaillard et al., 1999; and references therein). This subsidence was associated with the development of several depocenters, including the Manabí, Progreso, Gulf of Guayaquil-Tumbes, Zorritos, Talara-Sechura basins (Fig. 3.1B). The sedimentary rocks in these forearc basins are derived from relatively well-known sources such as the Cenozoic arc, Triassic granitoids, and Paleozoic metasedimentary rocks (Witt et al., 2023; Witt et al., 2017; this work). The proportion of zircon grains older than 200 Ma increases from north to south, ranging from approximately 10% of total grains in the Manabí basin to 45% in the Zorritos Basin (the onshore part of the Tumbes Basin). This trend, as well as other provenance characteristics, have been previously discussed by Witt et al. (2023).

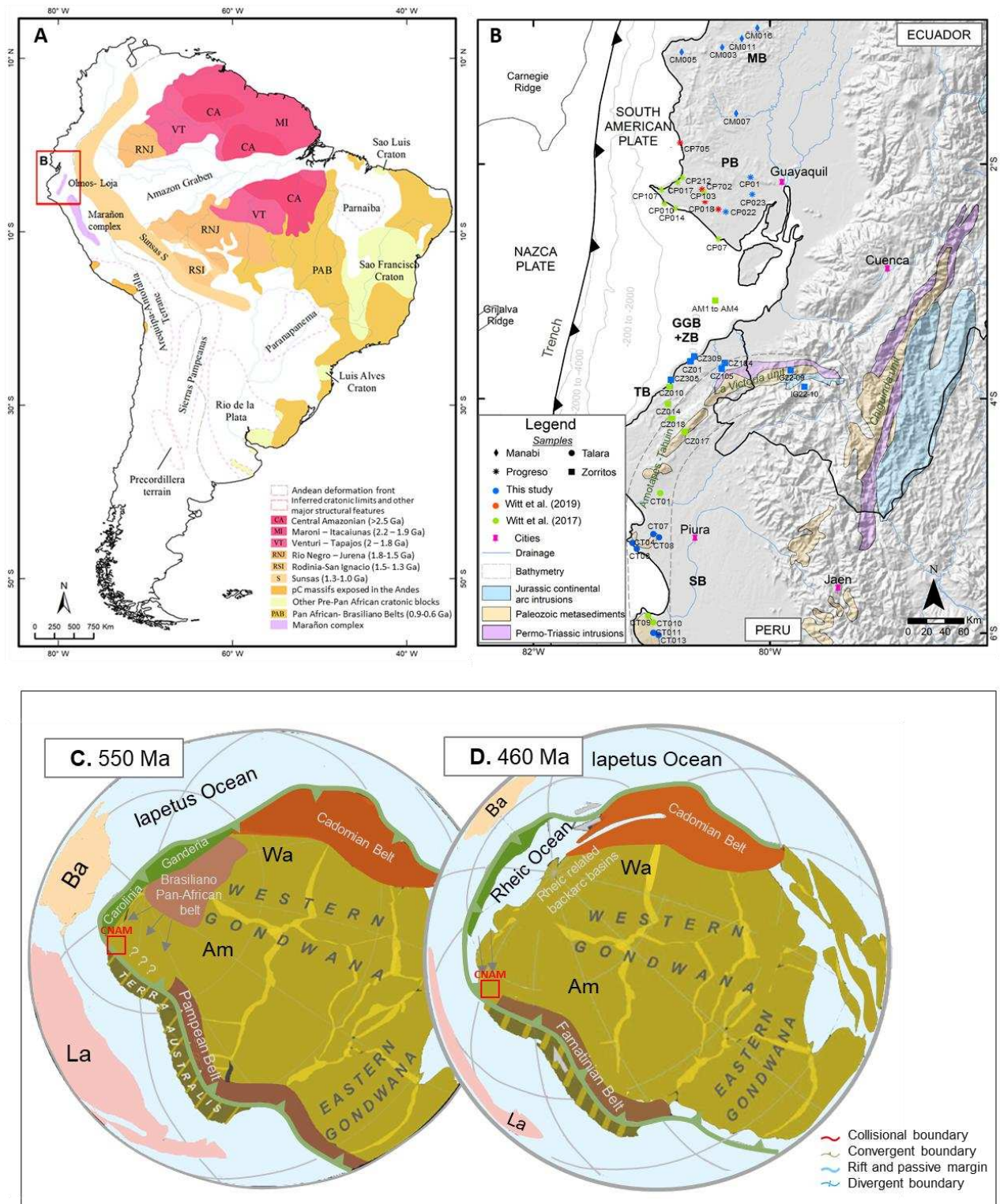


Figure 3.1 A) Map of the South American geochronological basement terranes (Modify from Ramos et al. 2009; Pepper et al. 2016). B) Geodynamic sketch of central-southern Ecuador and northern Peru. ATB: Amotapes-Tahuin Block; GGB+ZB: Gulf of Guayaquil and Zorritos Basin; MB: Manabí Basin; PB: Progreso Basin; SB: Sechura Basin; and TB: Talara Basin. (B-C) Simplified cartoon showing paleogeographic reconstructions from 550 Ma to 460 Ma, (modified from Oriolo et al. 2021). Abbreviations: Am—Amazonia; Ba—Baltica; La—Laurentia; WA—West Africa.

3.3 Sampling

We used 34 medium-grained sandstone samples, from Late Cretaceous to Pleistocene sedimentary outcrops across the Manabí, Progreso, Zorritos, Talara-Sechura Basins. These forearc deposits are complemented by late Cretaceous sedimentary rocks that outcrop in the Tortuga region. All sampling locations are presented in Fig. 3.1B and detailed in Table 2.1. We also analyzed ten samples from bedrock outcrops from the basement highs of the Amotape Complex as the Illescas massif—all sampling locations are once again presented in Table 2.1. These include seven fine- to medium-grained metaquartzites (CZ017, CT09, CT010, CT011, IG22-09, IG22-10, and IG22-43), and three coarse-grained granites (CZ104, CZ105, and CT013).

3.4 Analytical methods and results

3.4.1 Zircon geochronological data

A total of 15 detrital and eight bedrock samples were selected for zircon U-Pb dating using Laser Ablation ICPMS (La-ICP-MS). The sample codes and locations of these samples are presented in Fig. 3.1B and Table 2.1. A more detailed description of the analytical methods used is presented in chapter 2, section 2.3.

There were a total of 1311 U-Pb concordant dates obtained from the detrital and bedrock record, ranging from approximately 200 to 3000 Ma. Of these, 977 dates were obtained from detrital samples (Fig. 3.2A) and were distributed as follows: i) 425 dates obtained from the Progreso Basin, including 74 new dates; ii) 23 new dates from the Manabí Basin; iii) 208 previously published dates derived from the Amistad well; iv) 123 dates from the Zorritos Basin, including 37 new dates; and v) 198 dates from the Talara-Sechura basins, including 165 new dates.

This study reports the results of a total of 281 analyses conducted on metamorphic bedrock samples, including 62 analyses conducted on a low-grade metamorphic schist (CZ017; U-Pb data published in Witt et al. (2017)) and 219 grains from metaquartzites samples included both published (CT010) and new dates (CT09, CT011, IG22-09, IG22-10, and IG22-43). The maximum depositional ages (MDA) of the bedrock samples from the Talara-Sechura and Zorritos Basins were calculated using the methodology described by Dickinson and Gehrels (2009). Specifically, the $YC1\sigma(2+)$ age of each sample was determined: this refers to the mean age of the youngest two or more grains that overlap within a 2σ error range. In the Talara-Sechura Basins, the MDAs of the metasediments from samples CT09, CT010, and CT011 were 307.5 ± 4.9 Ma ($n = 2$), 330 ± 3.9 Ma ($n = 2$), and 342 ± 4.1 Ma ($n = 2$), respectively, all of which fall within the Carboniferous period. The MDAs of the samples in the Zorritos Basin, CZ017, IG22-09, and IG22-43, were 320 ± 3.5 Ma ($n = 3$), 319 ± 9 Ma ($n = 2$), and

312 ± 2.1 Ma ($n = 2$), respectively, which also occurred during the Carboniferous. The similar MDAs of the bedrocks from the Talara-Sechura and Zorritos Basins suggest that they were influenced by similar geological events, and most likely reflect the influence of a regional tectonic framework.

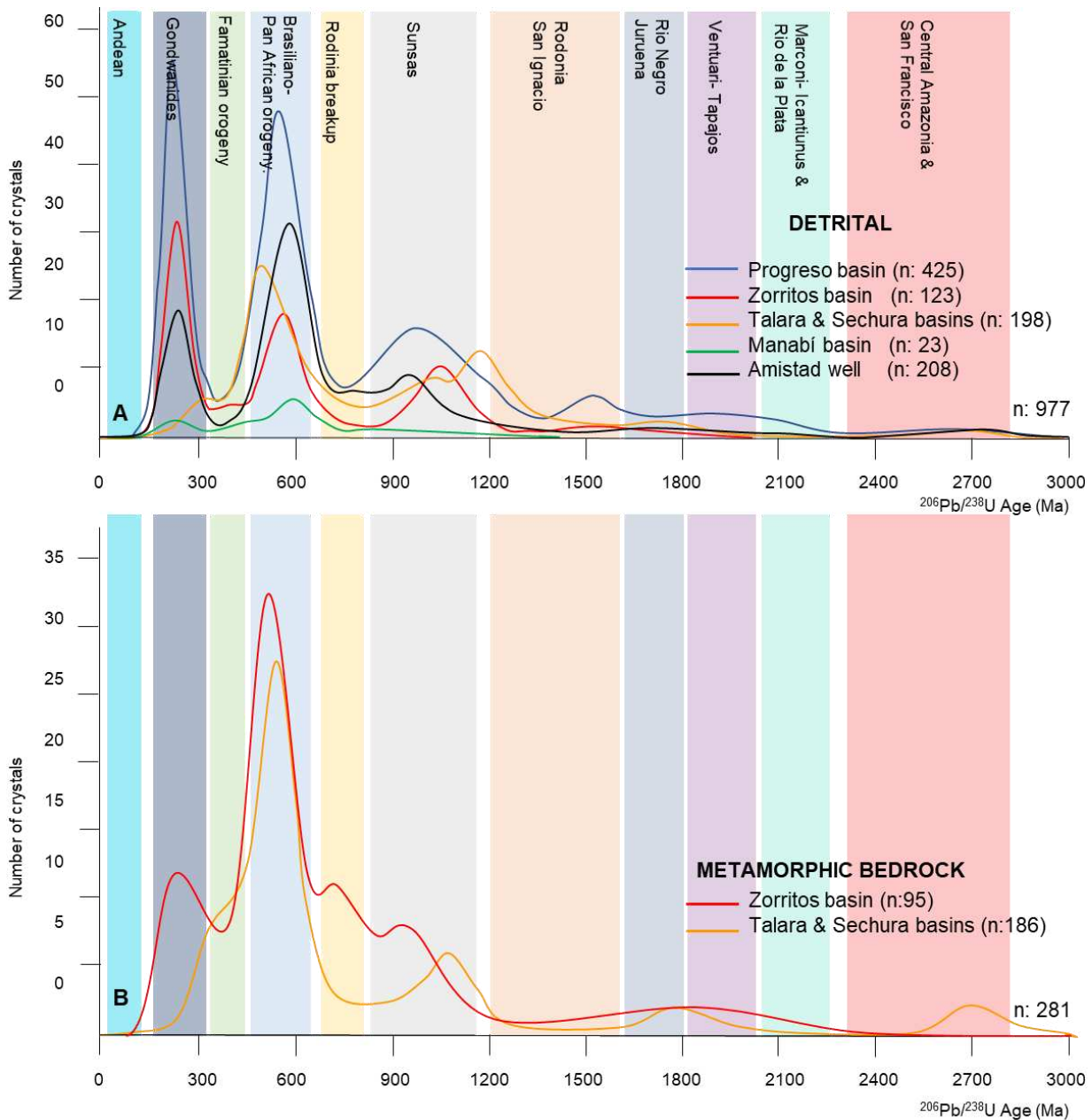


Figure 3.2 A) Zircon U-Pb ages probability distribution function of the Detrital record. B) Kolmogorov-Smirnov Multidimensional Scaling (MDS) plot assessing similarity between detrital datasets. C) Zircon U-Pb ages probability distribution function of the Metamorphic bedrock. Colors rectangles represent orogeny episodes and/or geological terranes discriminated in the Central-North Andean Margin (CNAM) reported at the literature.

The age distribution of the detrital zircons reveals a concentration around ages younger than 1000 Ma, comprising 74% of the total dataset. Two major populations are observed within this range (Fig. 3.3A): one clustered around ca. 640–460 Ma, with a peak around 560 Ma, and the other

concentrated around ca. 280–200 Ma, peaking at around 230 Ma. These were designated as Neoproterozoic–Ordovician and Permian–Triassic clusters, respectively. Two additional minor clusters were identified: one clustered around ca. 1000–890 Ma, peaking at around 960 Ma and another concentrated around ca. 390–310 Ma, with a peak at around 345 Ma; these clusters were defined as early Tonian and Late Devonian–Carboniferous, respectively. A distinct gap was also observed in the interval around 850–700 Ma, with only a few zircon ages recorded within this timeframe.

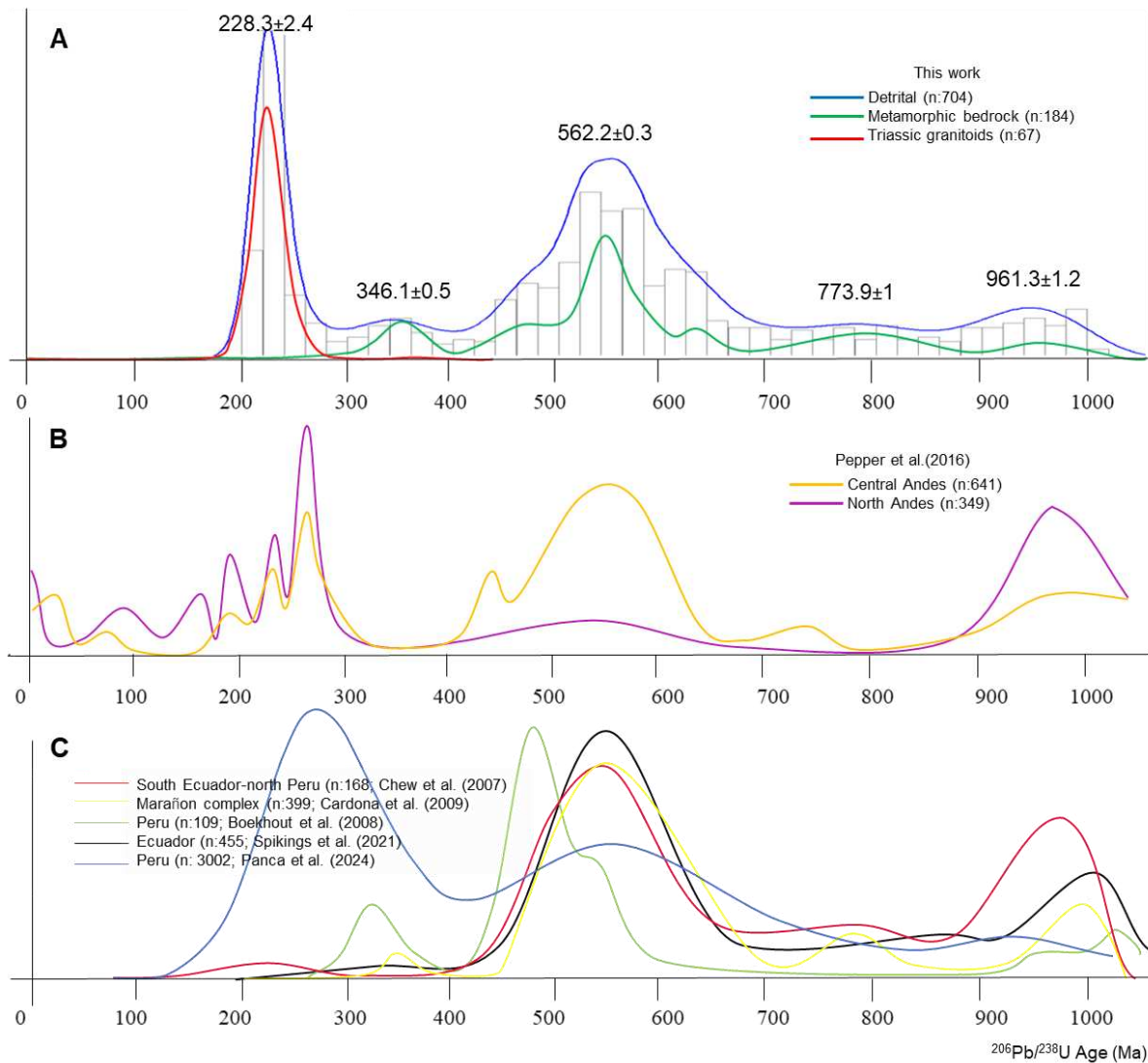


Figure 3.3 Zircon U-Pb age histograms and age probability distributions. Peaks for Kernel Density Estimation (KDE) were discriminated using ISOPLOT (Ludwig, 2003). A) This study. PDFs are shown for detrital, metamorphic and granitoids samples. B) Published data from Pepper et al. (2016). C) Published data from Chew et al. (2007) for southern Ecuador and northern Peru (red line); Cardona et al. (2009) for the Marañon complex (yellow line); Panca et al. (2024) for the Mitu Group in Peru (blue line); Spikings et al. (2021) for Ecuador (black line); and Boekhout et al. (2018) for the Ambo group in Peru (green line).

Results from 53 analyses were obtained from Triassic granitoids. The CZ104 granite yielded a $^{206}\text{Pb}/^{238}\text{U}$ mean date of 222 ± 1.2 Ma (Mean Square of Weighted Deviates (MSWD): 0.64; Fig.

3.4A); CZ105 exhibited a $^{206}\text{Pb}/^{238}\text{U}$ mean date of 229.2 ± 1.4 Ma (MSWD: 0.65; Fig. 3.4B); and a sample from the Illescas massif (CT013) exhibited a $^{206}\text{Pb}/^{238}\text{U}$ mean date of 222.8 ± 1.2 Ma (MSWD: 1.2; Fig. 3.4C).

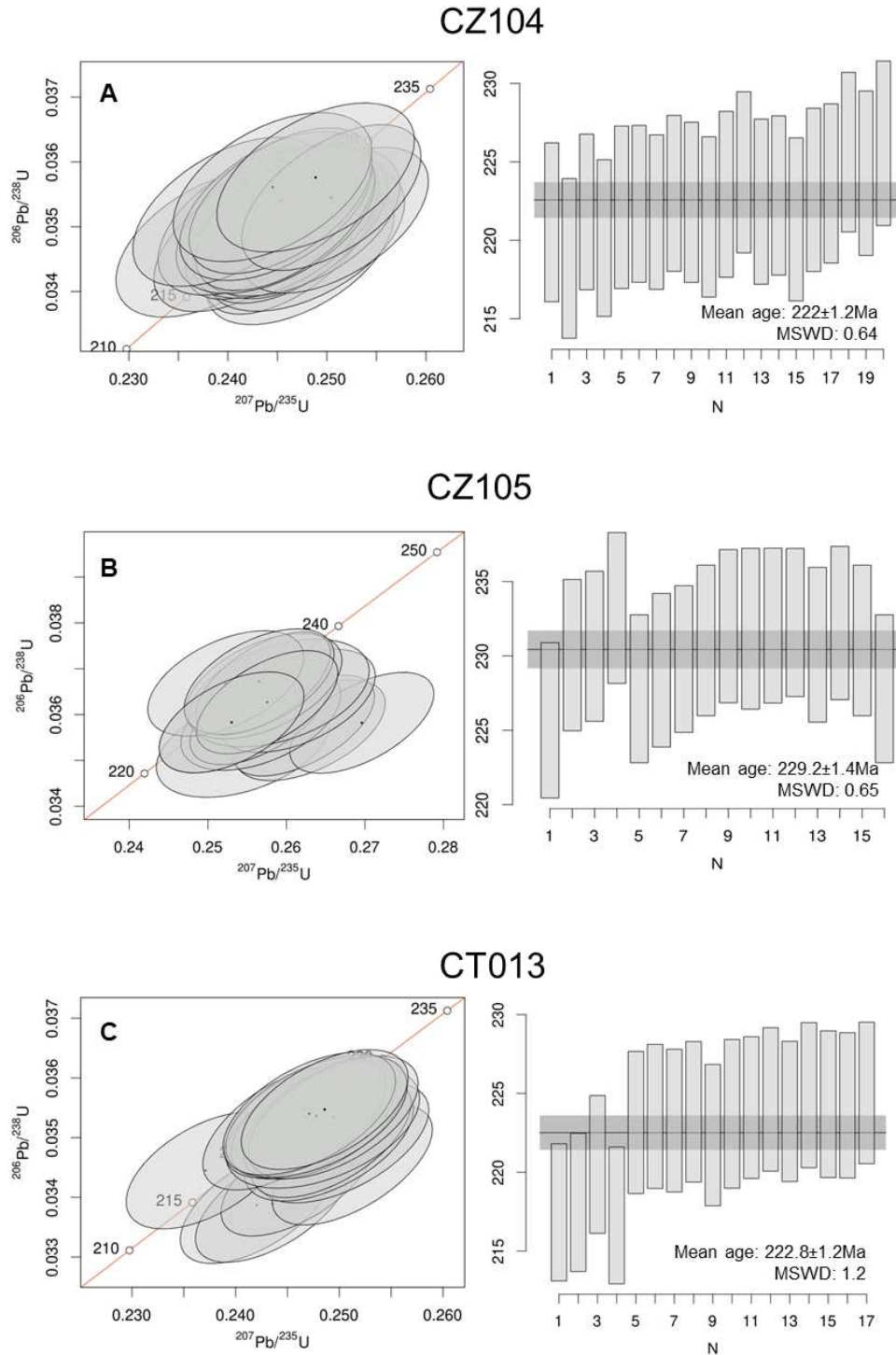


Figure 3.4 Concordia (left side) and weighted mean age (right side) diagrams for the studied Triassic granitoids. Errors are expressed at 2 sigmas. Plots were generated using ISOPLOT (Ludwig, 2003).

3.4.2 $\epsilon\text{Hf}_{(t)}$ analysis

A total of 175 $\epsilon\text{Hf}_{(t)}$ analyses were conducted on zircons from detrital samples (CZ018, CT004, and CT006), metamorphic bedrock (CT009, CT010, CT011, IG22-09, IG22-10, and IG22-43), and granite (CT013). The reported $\epsilon\text{Hf}_{(t)}$ values span a broad range from -39 to +16.3, with the majority exhibiting negative values (Fig. 3.5A). The overall trend of the data is somewhat ambiguous, with a general progression from depleted to enriched values from ~ 1000 Ma to ~ 600 Ma. This is followed by a more distinct trend that transitions from more highly enriched values ($\epsilon\text{Hf}_{(t)}$ up to -30) to less enriched values ($\epsilon\text{Hf}_{(t)}$ around 0) between ~ 600 Ma and ~ 300 Ma. In this latter interval, the most enriched signatures ($\epsilon\text{Hf}_{(t)}$ ranging from -36.5 to -28) are observed between ~ 560 Ma and 380 Ma. Finally, the period between ~ 300 Ma to 200 Ma period exhibits a more dispersed distribution, with $\epsilon\text{Hf}_{(t)}$ values fluctuating between -20 to 14.

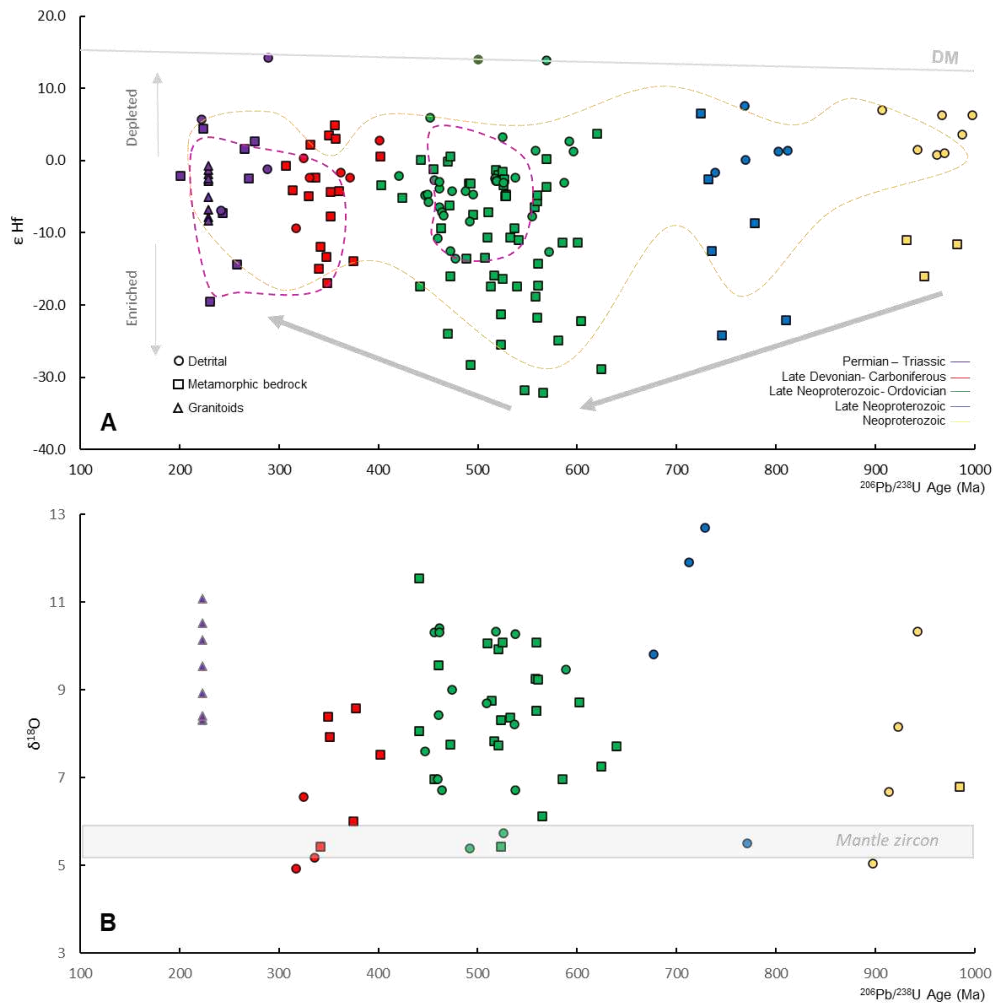


Figure 3.5 A) ϵHf plot versus age diagram. Dashed lines represent data from Pepper et al. (2016) and Capaldi et al. (2021), yellow and purple lines, respectively. B) $\delta^{18}\text{O}$ versus ages plot (mantle $\delta^{18}\text{O}$ boundaries are from Valley, 2003). Color

code represents different age clusters used also in figures 6, 7 and 8. Purple 200- 280Ma, red 310-390 Ma, green 450-640 Ma, blue 720-840 Ma & yellow 890-1010 Ma.

3.4.3 $\delta^{18}\text{O}$ analysis

A total of 80 grains from detrital sample CT006, metamorphic bedrock samples CT009 and CT010, and granite sample CT013 were subjected to $\delta^{18}\text{O}$ analysis. A majority of the zircon grains yielded supracrustal values that exceeded 7‰ (~93% of the data, Fig. 3.5B), while only a few grains exhibited $\delta^{18}\text{O}$ values typical of mantle reservoirs ($5.3 \pm 0.3\text{‰}$). Tonian to Cryogenian (1000 Ma to ~680 Ma) zircons possessed a wide range of $\delta^{18}\text{O}$ values, ranging between 5–12.7‰, with fewer than 15% exhibiting mantle-like signatures. The $\delta^{18}\text{O}$ values of zircons aged between 600–460 Ma were characteristic of crustal sources, with an average value of 8.4‰. In the subsequent ~380–300 Ma interval, the $\delta^{18}\text{O}$ values ranged from 4.6–8.5‰; this period was characterized by a notable increase in mantle-like signatures, which were observed in approximately 35% of the dataset during this period. Finally, the $\delta^{18}\text{O}$ values remained elevated (i.e., greater than 8‰) between ~300–200 Ma, with no values within the mantle range.

3.4.4 Trace element analysis in zircon

Trace element (TE) analyses were carried out on 167 zircon grains; the results of these analyses are presented chronologically following the five populations identified in Section 3.4.1. TE data were also filtered to avoid samples that were potentially affected by inclusions, following the criteria proposed by Attia et al. (2020). Specifically, samples with high counts of Y (>5000 ppm), and LREE ([Ce]> 200 ppm; [Pr]> 20 ppm; [Sm]> 300 ppm; [Eu]> 15 ppm and [Gd]> 150 ppm) were excluded from the analysis.

Tectono-magmatic provenance can be determined through the use of Hf/Th versus Th/Nb and U/Yb versus Nb/Yb ratios (Grimes et al., 2015). These ratios can distinguish between mid-ocean ridge, magmatic arc, and ocean island (as well as other plume-influenced) settings. Our dataset indicates that most of the studied zircon grains formed within a continental arc-related setting (Figs. 3.6A and 3.6B). This interpretation is further supported by their strong HREE enrichment patterns (Figs. 3.6C and 3.6D), pronounced positive Ce anomalies, and negative Eu anomalies; all of these observations are characteristic of arc-related environments (Figs. 3.6C and 3.6D). Notably, Late Devonian–Carboniferous grains (390–310 Ma) exhibited relatively low LREE depletion, while Permian–Triassic zircons exhibited higher HREE enrichment compared to other periods.

I-Type and S-Type granitoids can be distinguished through the use of Th, Pb, and Th/U versus Ce/U diagrams (Figs. 3.6E and 3.6F). I-type granitoids are characterized by lower Pb concentrations and higher Ce/U ratios, in contrast to S-type granitoids that exhibit elevated Pb levels and lower Th/U and Ce/U ratios (Roberts et al., 2024; Wang et al., 2012). Th/U ratios were used as an indicator of zircon petrogenesis and allowed for the distinction between metamorphic and igneous origins (Fig. 3.7A; Grimes et al., 2015). Zircon grains dated to ~1000–700 Ma and ~400–200 Ma exhibited values that were typical of magmatic zircon ($\text{Th/U} > 0.1$), while grains obtained from the Late Neoproterozoic–Ordovician cluster exhibited metamorphic signatures ($\text{Th/U} < 0.1$).

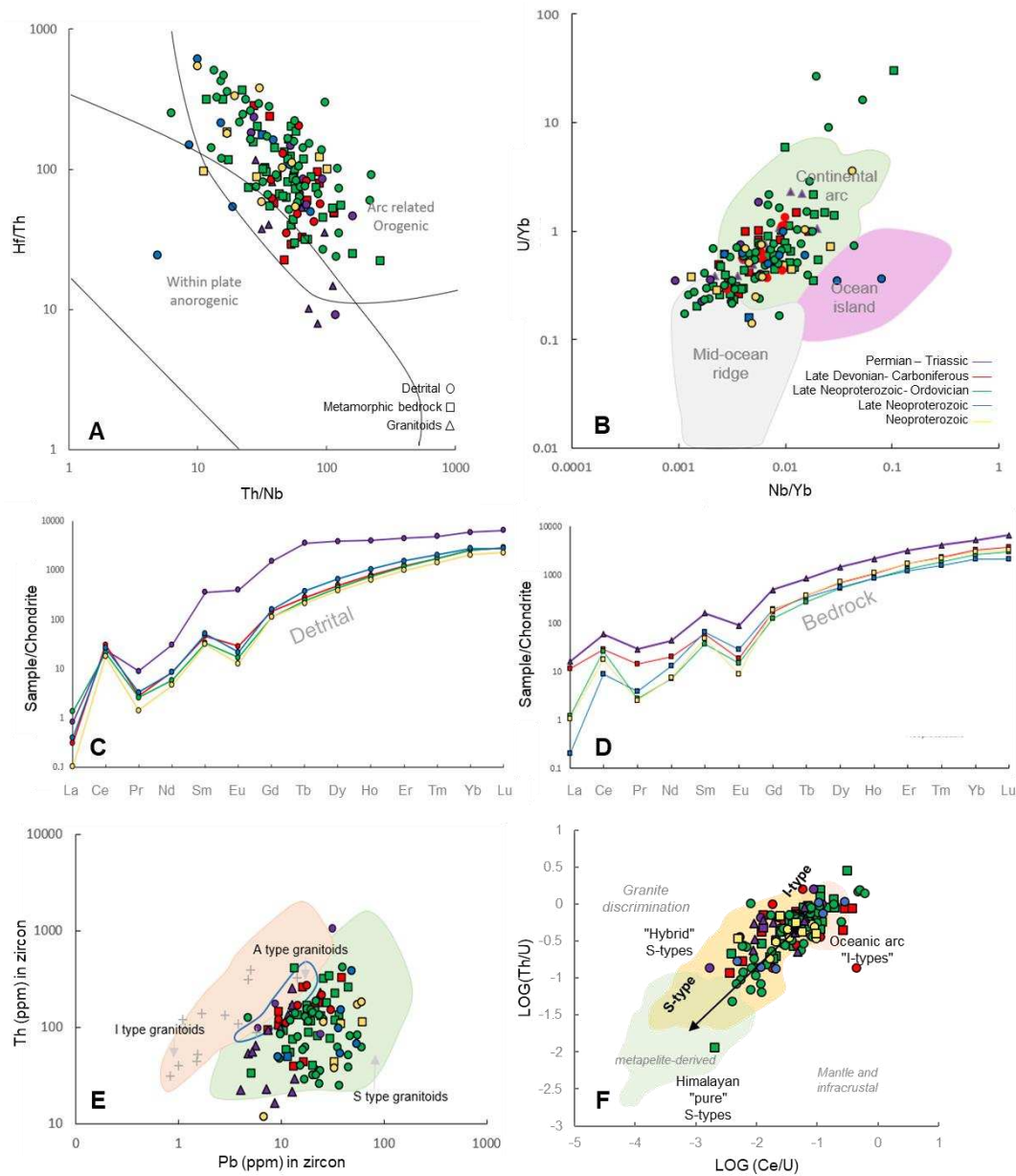


Figure 3.6 Zircon discrimination diagrams. (A-B) Hf/Th versus Th/Nb & U/Yb versus Nb/Yb (domains and contours are from Grimes et al., 2015 and Yang et al., 2012). (C-D) Spider diagrams displaying chondrite-normalized REE for detrital and metamorphic zircons. (E-F) I-Type and S-Type granitoids discriminants graphs (contours limits from Wang et al., 2012 and Roberts, et al., 2024). Gray crosses represent detrital zircon derived from the Cenozoic arc of southern Ecuador from Witt et al. (2023).

U/Yb ratio were used to distinguish between different magmatic sources and can also be used as a proxy of magmatic evolution (Fig. 3.7B; Grimes et al., 2015). The samples exhibited an increasing trend in U/Yb ratios from the Neoproterozoic to the Late Neoproterozoic-Ordovician, suggesting a greater contribution of crustal or enriched mantle materials. There was a subsequent slight decrease during the Late Devonian–Carboniferous, followed by another increase in the Permian–Triassic period.

LREE/HREE and La/Sm ratios can be used as a proxy to assess sediment and slab-derived fluid contributions to the magma source (i.e., Ikeda et al., 2016; Jara et al., 2021; Sellés et al., 2022). LREEs (including La) tend to be more incompatible during partial melting and magma differentiation compared to HREEs (including Sm). Consequently, La/Sm and LREE/HREE ratios increase as the crust becomes more enriched in LREE as a result of these processes (Luffi & Ducea, 2022). However, there were no discernible trends in the LREE/HREE and La/Sm ratios across the entire 1000 to 200 Ma timeframe (Fig. 3.7C and 3.7D).

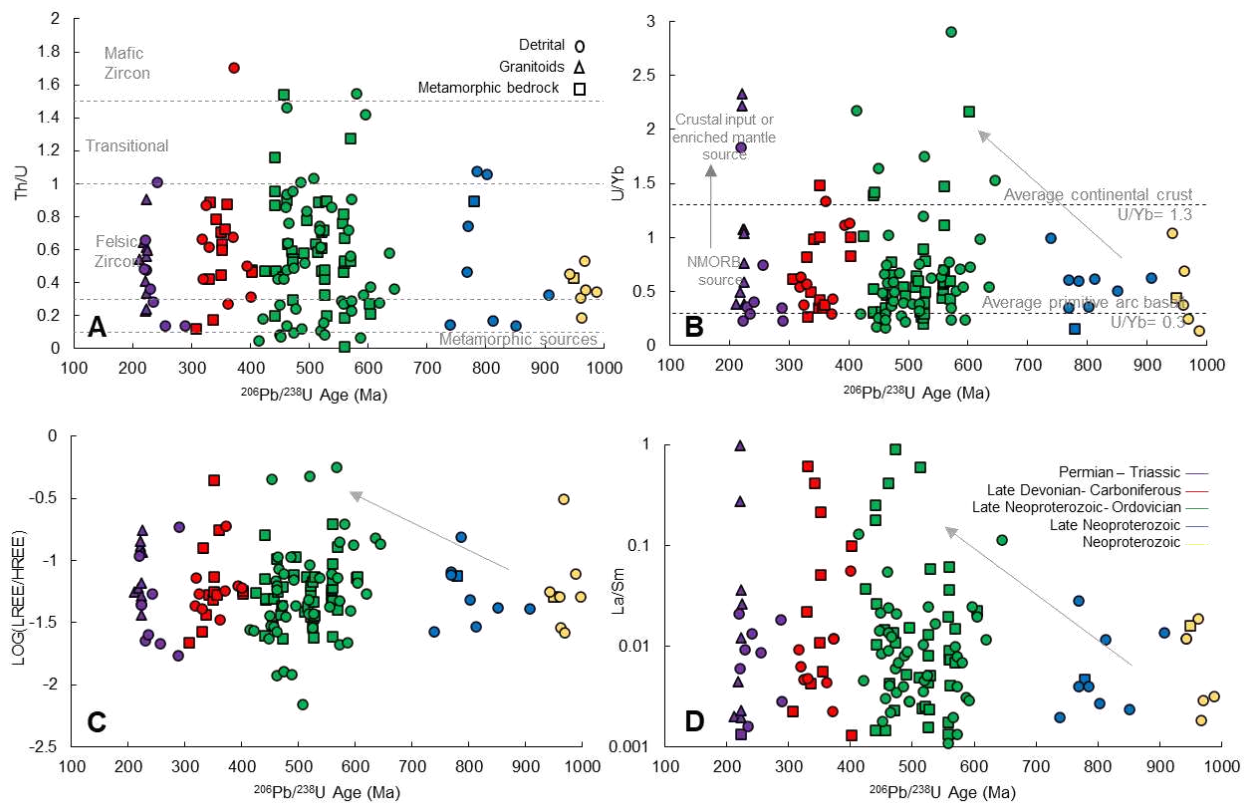


Figure 3.7 Geochemical ratios versus zircon U-Pb age plots. A) Th/U vs. U-Pb age (Ma), displaying the dominantly felsic nature of zircon grains. B) U/Yb vs. U-Pb age (Ma), exhibiting nature of zircon grains (domains and mean values are from Grimes et al., 2015). C) LOG(LREE/HREE); D) La/sm ratio. Where, LOG(LREE/HREE) and La/ Sm are proxies for sediment contributions to the magma source.

3.5 Discussion

3.5.1 Provenance model based on zircon chronology

To investigate the geological history of the 1000-200 Ma interval along the CNAM, we first assessed whether the forearc basins share a similar history. A classical Kolmogorov-Smirnov multidimensional scaling (MDS) analysis (Fig. 3.8) shows that the studied forearc basins have similar zircon age distributions. Among them, the Progreso, Amistad Well, Manabí, Talara-Sechura basins show a strong similarity, with Kolmogorov-Smirnov distances (KSD) of less than 0.2. In contrast, the Zorritos Basin shows slightly larger distances, up to 0.3, probably due to its higher proportion of 290-200 Ma zircons, which form the dominant population in this basin. This enrichment may be related to the proximity of the Triassic migmatites of the El Oro and Amotapes Massifs (e.g., Aspen et al., 1992; Riel et al., 2013; Riel et al., 2014; this study). Given the strong similarities between the datasets, we integrate the detrital zircon records for a collective analysis (Fig. 3.3A).

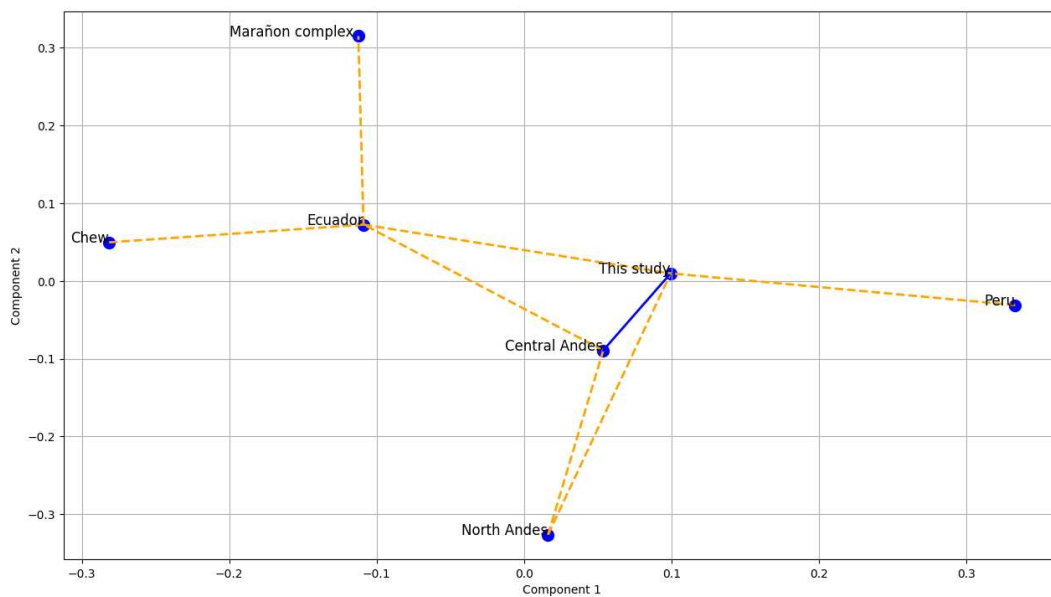


Figure 3.8 Multidimensional scaling (MDS) plot based on Kolmogorov-Smirnov (KS) distances for the studied forearc basins.

The geological history of the early Tonian (1000–890 Ma) interval is recorded in zircon grains from samples collected across both the south Ecuadorian and north Peruvian basins (Fig. 3.2A). The age populations of zircons in these basins generally cluster between 1000 and 890 Ma, with the exception of the Talara-Sechura Basins, where a higher proportion of older zircons forms an additional cluster

that peaks around 1100 Ma. In general, the Neoproterozoic event is consistent along the CNAM and is centered around ~960 Ma (Fig. 3.3A). This event, which is widely recognized across western Gondwana in several proto-Andean settings (Capaldi et al., 2021; Chew et al., 2007; Pepper et al., 2016; Spikings et al., 2021; Witt et al., 2017), is attributed to the Sunsás orogeny (Sunsás mobile belt 1.2 – 1.0 Ga; Santos et al., 2008 ; Fig. 1A) that resulted from the collision of the Arequipa (Antofalla) and Amazonia blocks during the assembly of Rodinia (Mišković et al., 2009; Pepper et al., 2016).

There were markedly fewer detrital zircons from the mid-Tonian to early Cryogenian (890–700 Ma), likely reflecting a substantial decline in magmatic activity and suggesting a potential gap in the magmatic history of the region (Fig. 3.3).

The magmatic history of the Ediacaran to the Middle Cambrian have been linked to the northern extension of the Pampean Orogeny (Chew et al., 2007; Spikings et al., 2021; Witt et al., 2017). Though there is a lack of direct evidence for a Pampean magmatic arc in Bolivia or southern Peru, Chew et al. (2008) proposed the existence of a buried late Neoproterozoic magmatic belt beneath the modern Andean belt or Amazon Basin based on detrital zircon data. They argued that eastern Amazonia was an unlikely sedimentary source due to the notable lack of detrital input from the core of the Amazon craton. Spikings et al. (2021), who also used data from detrital zircons, proposed that this signature could be related either to the magmatic arcs associated with the accretion of Pampia to western Gondwana (i.e., the Pampean Orogeny) or to contemporaneous orogenic belts that formed during the Brasiliano/Pan-African orogeny (now exposed in eastern South America). Due to the similar ages of the Pampean and Brasiliano/Pan-African orogenies, distinguishing the provenance of magmatic reservoirs based solely on zircon U-Pb ages without additional characterization is challenging (see Section 3.5.2).

Zircons from the Late Cambrian–Ordovician interval (510–460 Ma) can be more effectively contextualized within the framework of the opening of the Rheic Ocean along the northern margin of Gondwana (Fig. 3.1D; Nance et al., 2012) rather than the Famatinian orogeny. This interpretation is supported by the increasing prevalence of $\delta^{18}\text{O}$ mantle-like signatures and intermediate Hf values (-10 to +10; Fig. 3.5). Although Famatinian magmatism—recorded in the southern segments of South America (Bahlburg et al., 2020; Cardona et al., 2009; Chew et al., 2007)—represents a significant episode of regional magmatic activity, this period coincides with a gap in the zircon record in our samples. Consequently, we propose that Famatinian magmatism is unlikely to have been a major source for the sediments analyzed in this study.

The 380–310 Ma interval (which peaked at approximately 345 Ma) most likely corresponds to a Carboniferous magmatic event associated with arc magmatism in a sinistral stress regime, marking

the onset of the first Gondwanide orogeny during the assembly of Pangea (Boekhout et al., 2018; Panca et al., 2024). This magmatic episode is extensively documented in this region, especially across the northern and central segments of the Peruvian Eastern Cordillera (e.g., Balsas-Callangate and Pataz plutons). This event may be associated with the Mississippian Ambo Group (<359Ma; Cardona et al., 2009; Chew et al., 2007; Mišković et al., 2009), and is consistent with the MDAs reported in the Amotapes Massif and the Chiguinda–La Victoria units (Fig. 3.1B) by Witt et al. (2017) and Spikings et al. (2021), respectively.

Finally, the 270-200 Ma interval, peaking around 230 Ma, represents a significant phase of magmatism that can be associated with two different tectonic models. The first suggests a Triassic extensional setting, interpreted as a rift regime that facilitated seafloor spreading between Central and South America (e.g., Carrillo et al., 2021; Chew et al., 2016; Jaillard et al., 1995; Spikings et al., 2015). According to this model, the initial phase of rifting in western Gondwana was followed by a Late Triassic thermal sag associated with continued extension (Carrillo et al., 2021). Alternatively, Riel et al. (2018) propose that the Permian-Triassic evolution of western Gondwana was primarily driven by the reorganization of the Tethyan domain, with the associated magmatism attributed to the rotational dynamics of Pangea. This extensional event is recorded by the volcano-sedimentary Mitu Group in Peru (ca. 270-194 Ma; Chew et al., 2016; Mišković et al., 2009; Panca et al., 2024) as well as the Sabanilla, Moromoro, and Tres Lagunas units in Ecuador (Aspden et al., 1995; Aspen et al., 1992; Spikings et al., 2016; Villares et al., 2020). The abundance of detrital records around 220 Ma is consistent with the numerous exposures observed across southern Ecuador and north-central Peru (Chew et al., 2016; Cochrane et al., 2014; Gabriele, 2002; Riel et al., 2013; Riel et al., 2014; Villares et al., 2020; this work).

The U-Pb age spectra of zircon grains in this study (Fig. 3.3B) differ from data obtained from the northern Andes segment reported by Pepper et al. (2016); the main events in their study were Neoproterozoic and Late Paleozoic in age, and the 620–460 Ma cluster was categorized as a minor event. Instead, the results of this study are more consistent with the main events of the central Andes segment reported by Pepper et al. (2016), where the major (e.g., Neoproterozoic, Late Neoproterozoic–Ordovician, and Late Paleozoic) and minor events (Late Devonian–Carboniferous) are consistent with those described in Section 3.4.1 (Fig. 3.3B). Our dataset is also consistent with the detrital zircon age populations reported by Spikings et al. (2021) in the Ecuadorian and Peruvian segments (Fig. 3.3C), Panca et al. (2024) in Peruvian Mitu Group, Chew et al. (2007) in southern Ecuador and northern Peru, Cardona et al. (2009) in the Marañon complex (10°S; Fig. 3.3C), in which

significant zircon populations were identified in the Mesoproterozoic–Late Neoproterozoic, the Middle Neoproterozoic–Cambrian, and the Ordovician–Carboniferous.

3.5.2 Zircon petrochronology and reservoir characterization

The $\epsilon\text{Hf}_{(t)}$ and $\delta^{18}\text{O}$ isotopic systems are powerful tools for tracing magmatic sources and mantle dynamics in zircon records (i.e., Belousova et al., 2010; Dhuime et al., 2012; Kemp et al., 2006), and provide crucial insights into whether magmas are derived from the mantle, crustal materials, or a combination of both sources. For example, high positive $\epsilon\text{Hf}_{(t)}$ values are associated with depleted sources, whereas high negative values suggest an enriched component. Similarly, oxygen isotopes ($\delta^{18}\text{O}$) are instrumental in distinguishing between mantle-derived melts ($\delta^{18}\text{O} \approx 5.3\text{‰}$; Valley, 2003) and those influenced by crustal materials, which exhibit elevated $\delta^{18}\text{O}$ values (Peck et al., 2003; Valley et al., 2005). The isotopic signature of early Tonian materials is generally indicative of enriched sources ($\epsilon\text{Hf}_{(t)} = -16$ to $+6.3$; Fig. 3.5A); in combination with the $\delta^{18}\text{O}$ isotopic data, these results are indicative of a subordinate involvement of juvenile crustal and/or mantle-derived materials in the genesis of the magmas during this period ($\delta^{18}\text{O} = 6.68\text{--}10.32\text{‰}$; Fig. 3.5B).

A significant number of zircon grains dated to ca. 620–460 Ma exhibited highly variable $\epsilon\text{Hf}_{(t)}$ values ranging from -32.1 to $+13.9$ (though it should be noted that 93% of the $\epsilon\text{Hf}_{(t)}$ values were less than 0.5), and were associated with large $\delta^{18}\text{O}$ intervals ($5.38\text{--}11.54\text{‰}$). Oriolo et al. (2021) compiled a comprehensive $\epsilon\text{Hf}_{(t)}$ isotopic dataset to better understand the Early Paleozoic tectonic and geodynamic evolution of Western Gondwana. Their analysis revealed that rocks from several regions, including the Sierras Pampeanas, Patagonia, Puna, and the northern Chilean basement, formed between 650–440 Ma exhibited a uniform range of $\epsilon\text{Hf}_{(t)}$ values from -6 to $+1$. This consistent isotopic signature was interpreted as a characteristic feature of the Pampean Orogeny (550–510 Ma). However, our dataset reveals that zircon grains dated between 620–460 Ma exhibit $\epsilon\text{Hf}_{(t)}$ values ranging from -32.1 to $+13.9$. Specifically, $\epsilon\text{Hf}_{(t)}$ values from the time period between 620–510 Ma fluctuate between -32 and $+3$, in contrast with the previously reported homogeneous Pampean signal. Furthermore, the abundance of extremely negative $\epsilon\text{Hf}_{(t)}$ values (-30 to -20) indicates that these zircons came from a cratonic or ancient continental source (i.e., Chaudhuri, 2020; Diwu et al., 2024). Thus, we propose that Ediacaran to Middle Cambrian zircons found along the CNAM are derived from the Amazonian Craton and formed during the Brasiliano/Pan-African Orogeny (Fig. 3.1C). This widespread $\epsilon\text{Hf}_{(t)}$ signature has also been reported in Cadomian domains and is considered characteristic of Brasiliano/Pan-African magmatism (Oriolo et al., 2021).

Late Devonian–Carboniferous melts exhibit similarly variable $\epsilon\text{Hf}_{(t)}$ values ranging from -17 to $+3$; although this interval overlaps with intermediate values, they are still predominantly indicative of

enriched sources. Zircons from this period exhibit more consistent $\delta^{18}\text{O}$ values, with around one-third of grains falling within the mantle domain. These results suggest contributions from both enriched mantle and crustal sources. Additionally, Permian-Triassic zircon grains predominantly reflected enriched sources ($\epsilon\text{Hf}_{\text{(t)}}$ values ranging from -17 to +5.7) and exhibited elevated $\delta^{18}\text{O}$ values ($>8\%$) that were indicative of crustal maturation through extensive recycling processes and minimal juvenile input. This interpretation contrasts with findings published by Cochrane et al. (2014) and Spikings et al. (2015), who reported mostly high $\epsilon\text{Hf}_{\text{(t)}}$ values (+10 to +15) in metamorphic bedrock dated to 240–223 Ma. These values, which align closely with the depleted mantle array, led them to conclude that there was no significant continental crust assimilation after 225 Ma.

In general, zircon grains dated between 1000–200 Ma crystallized during repeated episodes of crustal reworking, as suggested by the highly enriched $\epsilon\text{Hf}_{\text{(t)}}$ values and the just few mantle-like $\delta^{18}\text{O}$ values. These periods of continental lithospheric reworking and enriched signatures have previously been associated with early phases of shortening and lithosphere thickening (Capaldi et al., 2021). However, Capaldi et al. (2021) identified only evolved contributions (i.e., $\epsilon\text{Hf}_{\text{(t)}}$ ranging from $\sim+4$ to -15) in the southern Central Andes from 550–200 Ma despite more juvenile contributions (mostly around 350 Ma) having been reported in the literature (i.e., Dahlquist et al., 2018; Hervé et al., 2014). Periods of evolved crustal processes have also been identified by Pepper et al. (2016), who identified three main trends of crustal evolution during the Sunsas, Brasiliano–Early Paleozoic, and Late Paleozoic events (~ 2000 –900 Ma, ~ 700 –400 Ma and ~ 350 –200 Ma, respectively), with mean $\epsilon\text{Hf}_{\text{(t)}}$ values ranging from +5 to -15. Consequently, our results indicate that the dominant process in the CNAM was crustal reworking rather than juvenile crustal addition. This suggests that some convergent margins, rather than being sites of net crustal growth, primarily function as zones of crustal differentiation over long timescales as shown in Eastern Australia, Western USA and the central Peruvian Andes (e.g., Belousova et al., 2010; Hawkesworth et al., 2019; Kemp et al., 2006).

Finally, the $\epsilon\text{Hf}_{\text{(t)}}$ isotopic data from our study are strongly consistent with central-eastern Gondwana signatures (i.e., central Africa, eastern South America), which are characterized by highly negative signatures (-20 to -40) that are indicative of partial melting of cratonic sources (i.e., George et al., 2015; Hueck et al., 2022; Milhomem Neto & Lafon, 2019). These values contrast with the less negative Andean signatures (i.e., Oriolo et al., 2021; Pepper et al., 2016). Based on these results, we propose that the Amazonian Camboriú Complex and the Punta del Este Terrane, situated in northeastern South America (see Hueck et al., 2022), represent the most probable sources of the highly evolved zircon grains analyzed in this study. The aforementioned terranes exhibit highly negative $\epsilon\text{Hf}_{\text{(t)}}$ values (-20 to -40) at 650–550 Ma, consistent with the patterns observed in the dataset.

Furthermore, the Camboriú Complex and Punta del Este terranes are characterized by relatively high populations of Paleoproterozoic and Neoproterozoic zircons (i.e., Gregory et al., 2015; Hueck et al., 2022; Martil, 2016; Vieira et al., 2019; Will et al., 2019), which coincide with prominent Ediacaran and minor Tonian peaks as well as the small but notable Paleoproterozoic population identified in this study (~1500 Ma in the detrital record of the Progreso Basin; Fig. 3.2). These findings suggest that the CNAM is more closely associated with the geological history of northern Gondwanan, specifically with the eastern Amazonian regions.

3.5.3 The zircon trace element record

Zircon geochemistry is a powerful tool for understanding the mechanisms that underpin crust-mantle interactions and the subsequent magma transport through the crust. Elements such as U, Th, Hf, Y, Yb, and Nb can be used to provide reliable information on the parent melt (Grimes et al., 2007; Grimes et al., 2015). When combined with U/Pb dating, the TE analysis of zircons is particularly well-suited to provide spatial and temporal insights into the magmatic versus metamorphic origin of zircon (i.e., U/Th), the geodynamic setting of related rocks (i.e., Th vs. Pb; U/Yb vs. Nb/Yb; Grimes et al., 2015), the genesis of S or I granites (i.e., Th vs Pb and Th/U vs Ce/U; Roberts et al., 2024; Wang et al., 2012), the enrichment of the parental source (i.e., U/Yb; Kotková et al., 2016), the degree of differentiation (i.e., U/Yb vs Hf; Griffin et al., 2006), and the crustal thickness evolution, among others (Luffi & Ducea, 2022, and references therein).

The analysis of zircons dating between 1000–200 Ma reveals that they crystallized in equilibrium with felsic melts (Fig. 3.7A), mainly in an orogenic arc environment with a continental signature (Figs. 3.6A and 3.6B). However, only a few within-plate anorogenic signatures were observed during the Late Neoproterozoic and Permian–Triassic events. Spider diagrams show the high HREE enrichment and positive Ce and negative Eu anomalies that are characteristic of rocks formed in arc-related settings (Figs. 3.6C and 3.6D).

U/Yb, LREE/HREE, and La/Sm ratios are commonly used in zircon geochemistry to distinguish between magmatic sources and processes. Our dataset exhibits a slight increase in these ratios at around 650 Ma (Figs. 3.7B, 3.7C, and 3.7D), indicating more enriched signatures during this period and suggesting the influence of cratonic sources in the magmatic reservoirs, most likely related to the Brasiliano/Pan-African orogeny.

Finally, our findings suggest that most zircons dating between 1000–200 Ma exhibit S-type or “hybrid” S-type signatures (Figs. 3.6E and 3.6F). This is consistent with Th vs. Pb plot and is partially consistent with Th/U vs. Ce/U plots. These results are also consistent with the results reported by

Aspen et al. (1992), who characterized the Triassic granitoids of Ecuador and Peru as pure S-type and linked them to the melting of low- to medium-grade semi-pelitic metamorphic rocks, a conclusion further confirmed by Riel et al. (2013); Riel et al. (2014). Based on these observations as well as the enriched isotopic nature of the analyzed zircons reported in this study, we hypothesize that a similar source may also account for older zircons dating back to ~650 Ma.

3.5 Conclusions

1. U-Pb ages obtained from zircons in our CNAM samples revealed two prominent populations: a Late Neoproterozoic-Ordovician population (640–460 Ma) and a Permian–Triassic population (280–200 Ma) that correspond to the Brasiliano-Pan-African and Gondwanides Orogenies, respectively.
2. Isotopic data suggest that there were continuous episodes of crustal reworking, primarily involving enriched material with only sporadic contributions from the mantle (only 7% of the $\delta^{18}\text{O}$ data exhibited values within $5.3 \pm 0.6\text{‰}$).
3. The most enriched zircon grains occur around 650 Ma, reflecting the influence of the Brasiliano/Pan-African orogeny during this period.
4. The U-Pb age spectra for this section of the Northern Andes (central-south Ecuador and northern Peru) are relatively similar to that of the central Andean region. However, the broader isotopic data suggest that the CNAM is more closely aligned with Amazonian signatures, indicating a stronger affinity with northern Gondwana rather than the typical Andean characteristics.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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CHAPTER 4: MESOZOIC (240 TO 90 MA) MAGMATISM IN THE ANDES OF SOUTHERN ECUADOR – NORTHERN PERU: GEODYNAMIC INSIGHTS FROM WHOLE-ROCK CHEMISTRY AND ZIRCON PETROCHRONOLOGY

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Declaration statement

This manuscript is under review in Lithos.

CRedit authorship contribution statement

Sandoval Espinel, L.C.: Conceptualization, Investigation, Visualization, Data curation, Methodology, Writing – original draft. **Witt, C.:** Conceptualization, Supervision, Funding, Writing – review & editing. **Iglesias J.:** Methodology, Writing – review & editing. **Zattin, M.:** Conceptualization, Supervision, Funding, Writing – review & editing. **Bruguier O.:** Methodology, Data curation, Writing – review & editing. **Poujol, M.:** Methodology, Data curation, Writing – review & editing. **Samaniego Pablo:** Writing – review & editing. **Bermudez, M.:** Funding, Writing – review & editing. **Sandoval Espinel, J.J.:** Visualization, Writing – review & editing.

HIGHLIGHTS

1. Triassic granitoids show strong crustal assimilation at least until 220 Ma.
2. New Jurassic ages challenge the regional south-younging magmatic trend and the missing Jurassic arc in northern Peru.
3. Magmatism migrated westward since middle Cretaceous ca. 110 Ma, later than in previous models.
4. The $\varepsilon H(t)$ and $\delta^{18}O$ data show a clear temporal trend from ~220 Ma to ~90 Ma towards more depleted mantle-like sources.

ABSTRACT

Southern Ecuador - Northern Peru encompasses a region where the Andean Mountain range shifts its orientation from N20°E in the north to N20°W in the south, marking the transition between the northern and central Andes, along the Huancabamba deflection. This study reconstructs the Mesozoic magmatic history of this key region by integrating petrography, U-Pb geochronology, whole-rock geochemistry and zircon trace-elements, $\epsilon\text{Hf}_{(t)}$ and $\delta^{18}\text{O}$. Our results suggest that much of the Mesozoic magmatism likely occurred in an extensional arc setting, with magmas progressively derived from increasingly depleted mantle sources, while crustal contributions diminished over time and crustal thickness remained relatively stable, at around 35–30 km. Magmas evolved both spatially and temporally, beginning with a wide Triassic arc dominated by granitoids exhibiting strong crustal signatures at least until 220 Ma. This was followed by mildly enriched signatures associated with a Jurassic to Late Cretaceous arc that remained stationary and active between ~190 and ~125 Ma. Around 110 Ma, the arc underwent a significant westward migration, potentially driven by slab rollback, coinciding with the opening of the Celica-Lancones Basin. This migration is consistent with westward magmatic shifts documented in central Ecuador, Colombia, and northern Peru. In contrast, this westward migration differs from the coeval eastward magmatic migration observed in central and southern Peru. Additionally, new U-Pb ages challenge the notion of a missing Jurassic arc in northern Peru, providing evidence of Jurassic magmatism extending southward to at least 6°S.

Key-words: Zircon, U-Pb dating, E-MORB crust, magmatic reservoirs, geochemistry, Andes.

4.1 Introduction

Magmatism at convergent margins is primarily controlled by subduction dynamics, which influences mantle flow patterns and governs associated tectonic processes. A key factor is the variability in slab dip, which affects mantle flow within the wedge, melt generation, and the thermal structure of the overriding lithosphere (Capitanio et al., 2010; Ducea et al., 2015). These parameters, in turn, strongly influence the position of magmatic arcs and the emplacement of plutons and volcanic centers. For instance, a steeper slab dip typically promotes the ascent of hotter mantle material into the wedge corner, shifting magmatic arcs closer to the trench (e.g., Ha et al., 2023).

One of the most prominent examples of a convergent margin in a continental context is the Andean orogenic belt, which developed above an eastward-subducting system. Here, the oceanic Nazca Plate (Fig. 4.1; formerly part of the Farallon Plate) is being subducted along much of the South American continental margin (DeCelles et al., 2009; Horton, 2018). Within this framework, the Northern Andes have undergone a complex history of magmatism from the Paleozoic to the Cenozoic (Leal-Mejía et

al., 2019; Mišković et al., 2009; Witt et al., 2023). Despite recent advances in understanding the long-term magmatic and tectonic evolution of the region (i.e., Boekhout et al., 2012; Carrasco et al., 2023; Chavarría et al., 2022; Cochrane et al., 2014; Duan et al., 2022; Hughes & Pilatasig, 2002; Jaillard, 2022; Jaillard et al., 2005; Pratt et al., 2005; Riel et al., 2014; Sandoval-Espinel et al., 2025; Spikings et al., 2000; Vallejo et al., 2009; Winter et al., 2010; Witt et al., 2017), significant uncertainties remain regarding the timing and processes that shaped this segment of the orogen.

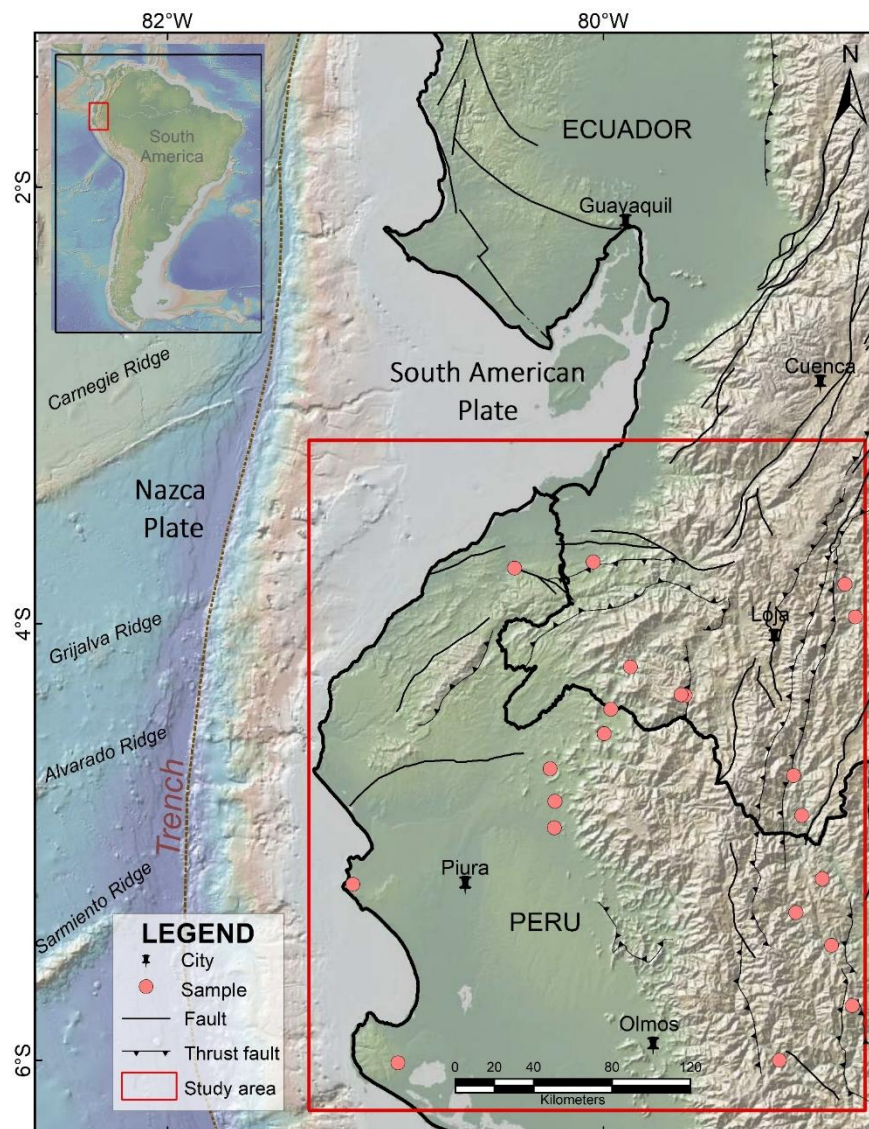


Figure 4.1 Geodynamic sketch of the southernmost Northern Andes, showing the location of samples from this study. Topographic and bathymetric data were extracted from GeoMapApp (<http://www.geomapapp.org>).

Tectonic models from the southernmost northern Andes offer contrasting interpretations (see Bustamante et al., 2016; Cochrane et al., 2014; Spikings et al., 2015), partly due to the limited availability of geochronological data. Most existing geochronological constraints from southern Ecuador and northern Peru are derived from Ar-Ar and K-Ar dating (e.g., Baldock, 1977; Cochrane et al., 2014; Feininger & Silberman, 1982; Hama, 1990; Litherland, 1994; Spikings et al., 2010;

Winter, 2008). In contrast, U-Pb zircon ages, which may in most-cases provide more robust geochronological constraints, remain scarce (e.g., CGS_INGEMMET, 2017; Cochrane et al., 2014; Noble et al., 1997; Winter, 2008), leaving open questions about the temporal and geodynamic evolution of the area.

In this study, we integrate U-Pb zircon geochronology, petrographic analyses, and whole-rock and zircon geochemistry from intrusive rocks exposed in the southernmost northern Andes (southern Ecuador-northern Peru; $\sim 4^{\circ}\text{S}$ to $\sim 6^{\circ}\text{S}$). These new constraints allow us to significantly refine the character and spatio-temporal evolution of Mesozoic magmatism between 240 and 90 Ma, to critically reassess the timing and mechanisms of the Cretaceous westward arc migration in this segment of the Andes, and to advance more robust, updated geodynamic models.

4.2 Geological context

The northern Andes span from Venezuela to northern Peru, and are separated from the central Andes by the Huancabamba Deflection at $\sim 5^{\circ}\text{S}$, marking a change in the orientation of the Andean belt from $\text{N}20^{\circ}\text{E}$ in the north to $\text{N}20^{\circ}\text{W}$ in the south (Fig.4.1). The magmatic activity along the northern Andes has been the product of the interaction of the South American plate with a series of oceanic plate systems, resulting in discontinuous magmatic events that took place from Carboniferous to Holocene, in a long-lived subduction setting (Horton, 2018; Jaillard et al., 2000).

During the early Mesozoic, the continental margin underwent an extensional regime that triggered extensive magmatic activity between ~ 265 and ~ 215 Ma. This magmatism, commonly referred to as the Choiyoi event, resulted from prolonged high mantle heat flow and crustal anatexis, (e.g., Cochrane et al., 2014; Riel et al., 2018). Although various models have been proposed to explain this Triassic event, they can be broadly categorized into two groups: 1) One group of studies argues that magmatism occurred in the absence of subduction, attributing the extensional tectonics to rifting and the progressive breakup of Pangea (Fig. 4.2.; Carrillo et al., 2021; Chew et al., 2016; Cochrane et al., 2014; Spikings et al., 2015). 2) Another group proposes that subduction remained active during this time, with processes such as slab steepening and rollback driving both the extension and associated magmatism (e.g., Riel et al., 2018; Villares et al., 2020). Independently of the model, crustal anatexis during this period resulted in the formation of S-type Triassic granitoids, which are currently present along the Eastern Cordillera and the Amotape – Tahuin block (ATB) in Ecuador (e.g., Aspen et al., 1992; Litherland, 1994; Riel et al., 2014; Sandoval-Espinel et al., 2025; Spikings et al., 2015) and in the southern and central segments of the Peruvian Eastern cordillera (Mišković et al., 2009). Following this extensional event, it is believed that the Pacific margin remained passive until approximately ~ 209 Ma (Gómez et al., 2021; Pindell & Erikson, 1994).

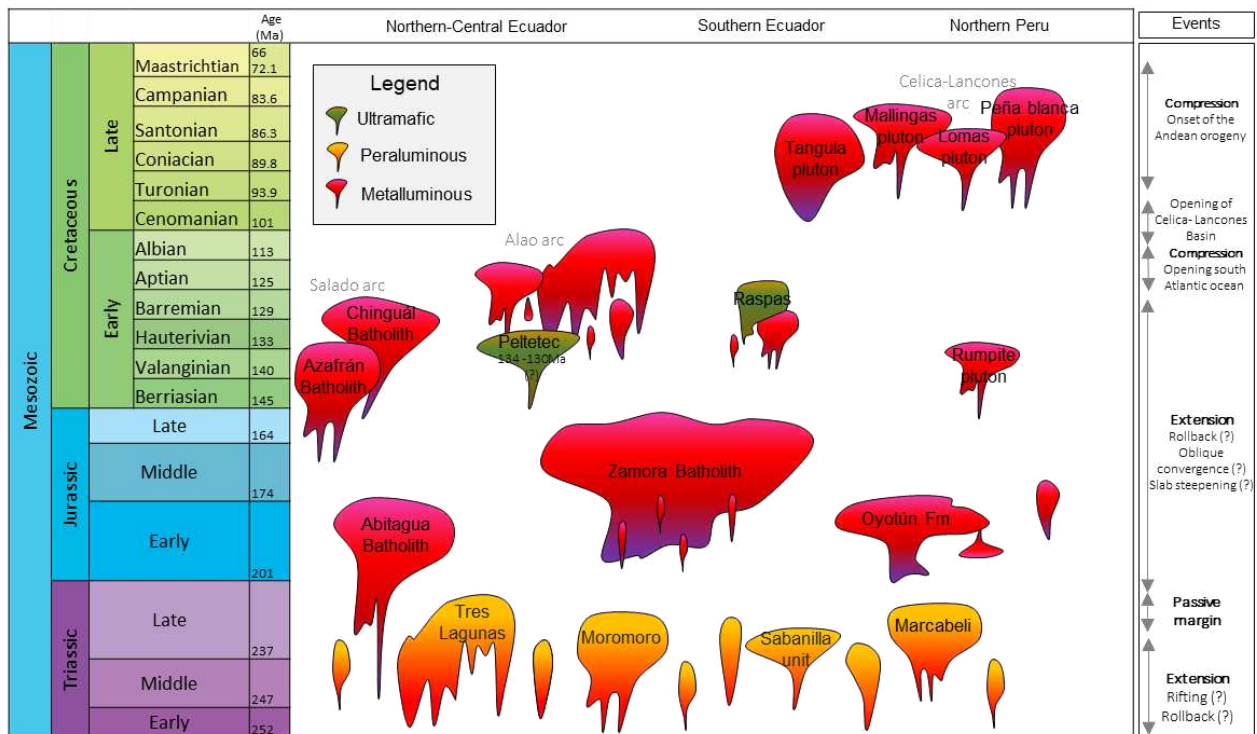


Figure 4.2 Overview of the spatiotemporal distribution of Mesozoic magmatism in southern Ecuador and northern Peru, along with key tectonic and geodynamic events. The figure is based on data from CGS-INGEMMET (2017), Cochrane et al. (2014), Feininger and Silberman (1982), Hama (1990), Litherland et al. (1994), Noble et al. (1997), Schutte et al. (2010), Spikings et al. (2015), Ulrich (2005), Winter (2008), Sandoval et al. (2025), and this study.

The reestablishment of subduction occurred between ~209 and ~190 Ma (Cochrane et al., 2014), marking the onset of a new continental arc inboard of South America (Zapata et al., 2016). During the Jurassic, the northern Andes experienced an extensional regime, which has been attributed to either: (i) highly oblique convergence, resulting in significant sinistral strike-slip motion along the northern Andes (Bustamante et al., 2016; Jaillard et al., 1990) ; or (ii) slab rollback, which drove the westward migration of the arc between 189 and 141 Ma, from the present-day Cordillera Real to the Western Cordillera (Cochrane et al., 2014; Spikings et al., 2015). The Jurassic arc extends across much of the northern Andes, with older crystallization ages (e.g., 202-196 Ma) recorded in northern Colombia (e.g., Santander and Santa Marta massifs; Leal-Mejía et al., 2019; Van der Lelij, 2013; Villagómez et al., 2011) and younger ages (e.g., 174-132 Ma) reported in Ecuador (e.g., Abitagua and Zamora batholiths; Chiaradia, 2009; Cochrane et al., 2014). This distribution has been initially interpreted as a general southward-younging magmatic trend from northern Colombia to southern Ecuador (e.g., Spikings et al., 2015). Furthermore, Jurassic magmatism is thought to be absent between 5°S and 12°S, either because the arc never formed or was removed by subduction erosion (Boekhout et al., 2012; Spikings et al., 2015).

Extensional conditions persisted until the Early Cretaceous, facilitating the emplacement of the Peruvian Coastal Batholith within the Western Cordillera of Peru (Boekhout et al., 2012; Cochrane

et al., 2014; Mišković et al., 2009), as well as the formation of the Salado and Alao arcs in Ecuador (Carrasco et al., 2023). The Salado Arc, located in northern and central Ecuador, consists of metaluminous granitoids, such as the Azafrán Batholith (~145–140 Ma; Cochrane et al., 2014) and the Chingual Pluton (~125 Ma; Spikings et al., 2015). Westward, the Alao Arc is composed of two distinct rock assemblages: (1) Calc-alkaline volcanic rocks, formed between ~140 and 110 Ma (Carrasco et al., 2023) and (2) E-MORB-like basalts comprising the Peltetec Unit (130–134 Ma; Gabriele, 2002; Spikings et al., 2015). Notably, in southern Ecuador, rocks with lithological similarities to these E-MORB-like basalts have been identified within the ophiolitic succession of the Raspas Complex (130–120 Ma), which has undergone metamorphism under blueschist to eclogite facies conditions (Gabriele, 2002; Riel et al., 2014). Nevertheless, recent U-Pb dating suggests that some units previously identified as Early Cretaceous in central Ecuador (west of the older arcs), such as the Peltetec Complex, may instead be Triassic in age (~228 Ma; Villares et al., 2020).

Compression may have started around 120–110 Ma, likely in response to increased convergence rates associated with the opening of the South Atlantic Ocean (Eagles, 2007; Maloney et al., 2013). This compressive phase is thought to be followed by trench-parallel extension between ~110 and 100 Ma, resulting in the opening of the NE-SW trending Celica-Lancones basin between the Early Cretaceous volcanic arc and the ATB (Jaillard et al., 1999; Mourier, 1988; Winter et al., 2010).

Around 100 Ma, active seafloor spreading in the South Atlantic and the resulting westward drift of the South American Plate (Eagles, 2007) triggered a tectonic shift from extension to compression. This transition led to the development of the Early to Late Cretaceous Celica-Lancones Arc (~110–78 Ma; Schütte et al., 2010; Soler, 1991; Winter, 2008), which extends across southern Ecuador and northern Peru (Fig. 4.3). The arc is composed mainly of andesitic lavas, pyroclastic deposits, and intrusive bodies, including the Tangula pluton in Ecuador and the Peña Blanca, Lomas, and Mallingas plutons in Peru (e.g., Schütte et al., 2010; Winter, 2008). Owing to limited geochemical and geochronological constraints, some authors have proposed that the Celica–Lancones Arc may be genetically linked to the same magmatic pulse responsible for the formation of the Alao Arc (Jaillard et al., 1999; Spikings et al., 2015). Farther south, in central and south Peru, the compressional regime is recorded by localized folding and faulting, the cessation of marine sedimentation and widespread emplacement of large intrusive bodies (Martínez-Ardila et al., 2023), contributing to the final stages of the formation of the Peruvian Coastal Batholith (Cobbing & Pitcher, 1983).

4.3 Results

Nineteen igneous rock samples were collected from outcrops located between 3.5°S and 6°S in southern Ecuador and northern Peru. These samples were obtained from the Zamora Batholith, the

Celica-Lancones arc, the ATB, the Rumpite Pluton, the Tauromarca and the Chunchuca-Olimpo units (for the nomenclature of the latter two units, see Jaimes Salcedo et al. (2013) and Sanchez et al. (1996)). Sample locations and corresponding nomenclature are provided in Table 2.1 and illustrated in Figure 4.3. Whole-rock analytical procedures, including petrographic descriptions and major and trace element geochemistry by ICP-AES, are detailed in chapter 2.

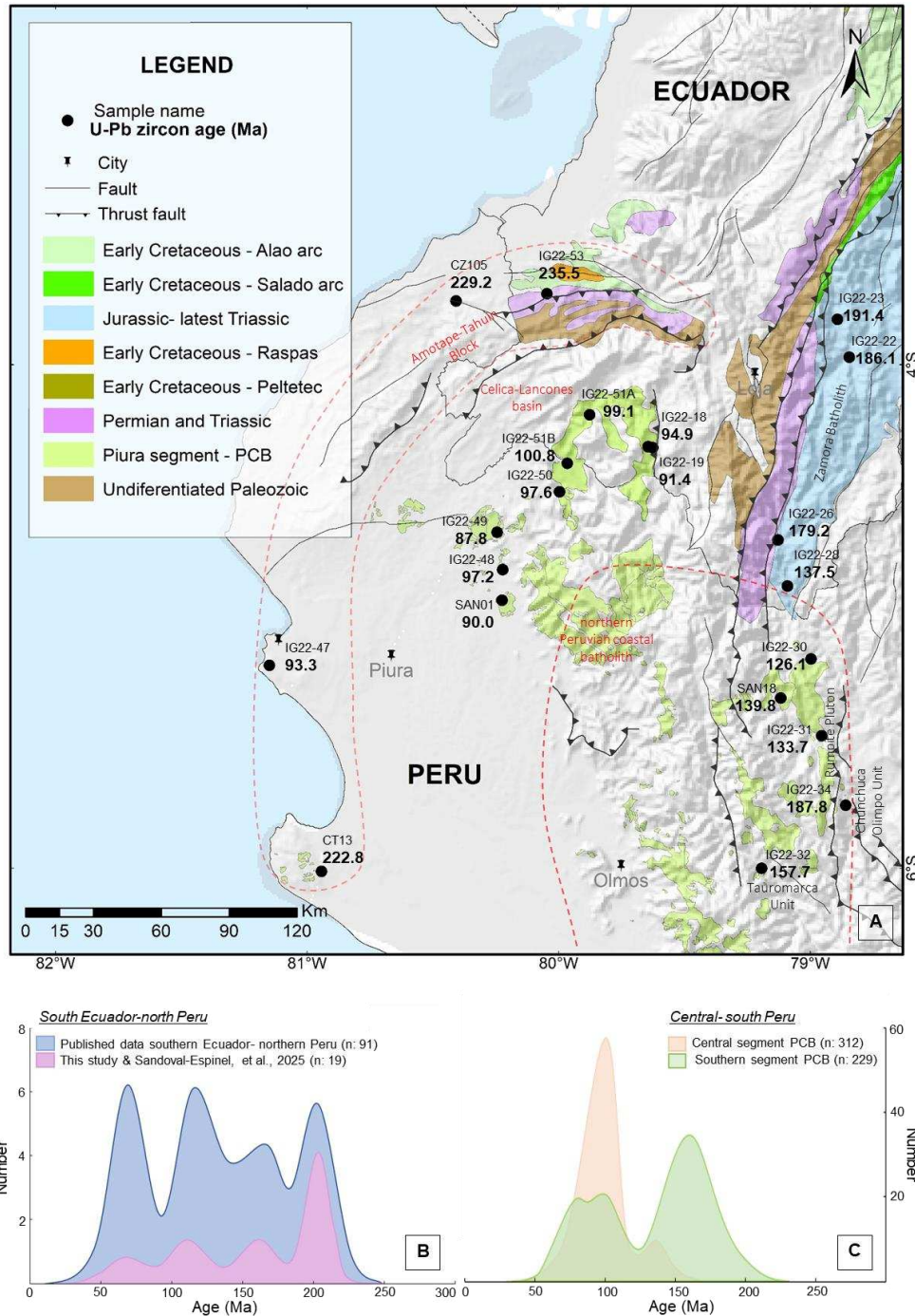
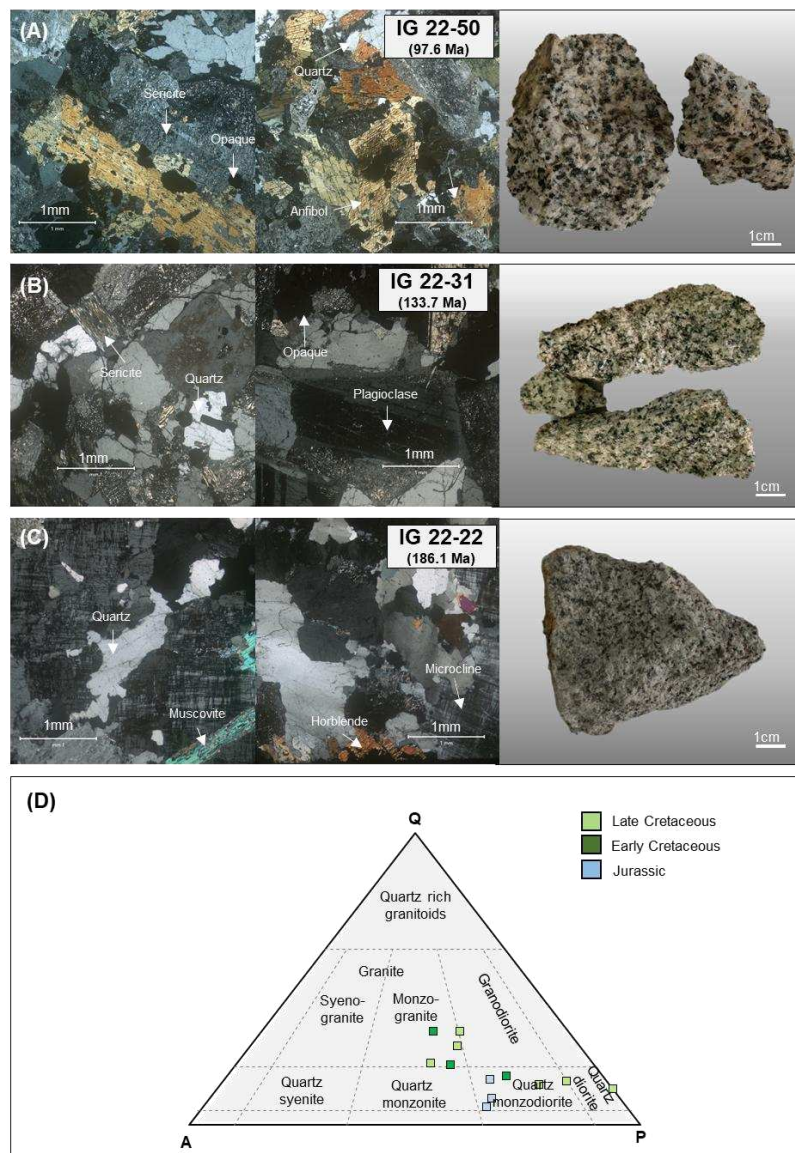


Figure 4.3 (A) Map of the study area showing sample locations and their corresponding U-Pb zircon ages. The geological background represents Ordovician to Cretaceous rocks. (B) Data from the northern segment (3 – 6°S). (C) Data reported by Martínez-Ardila et al., (2023) for the central and southern segments. Data represent in situ rocks based on age, ranging

from 66 Ma to 250 Ma. Age spectra were generated using IsoplotR, with kernel bandwidth set at 10 Ma for all graphs. PCB: Peruvian Coastal Batholith.

4.3.1 Petrographic analyses

Petrographic characterization was performed on twelve thin sections, with 200-point counts conducted on each. All samples display phaneritic textures (Figs. 4.4), with crystal sizes ranging from 0.8 to 2.7 mm (Fig. 4.4). The primary mineral assemblage includes quartz, orthoclase, microcline, plagioclase, muscovite and biotite phenocrysts, accompanied by minor amounts of hornblende, apatite and opaque minerals (Table 2.3). The rocks are generally fresh, with well-preserved grain boundaries and limited evidence of alteration. However, minor sericitization of plagioclase and slight chloritization of biotite are observed locally.



The QAP diagram (Streckeisen, 1974) was used to classify the rocks according to their mineralogical or modal composition. The analyzed rocks are predominately acidic in composition. Jurassic samples plot in the quartz monzodiorite field whereas Early Cretaceous rocks fall into the monzogranite and quartz monzodiorite fields. In contrast Late Cretaceous rocks display a wider compositional range, including quartz diorite, quartz monzodiorite and monzogranite (Fig. 4.4D).

4.3.2 Zircon U-Pb geochronology

We present nineteen new zircon U-Pb ages, along with two Triassic ages previously reported by Sandoval-Espinel et al. (2025): samples CT013 (222.8 ± 1.2 Ma) and CZ105 (229.2 ± 1.4 Ma). Selected concordia diagrams and weighted mean age plots are presented in Fig. 4.5.

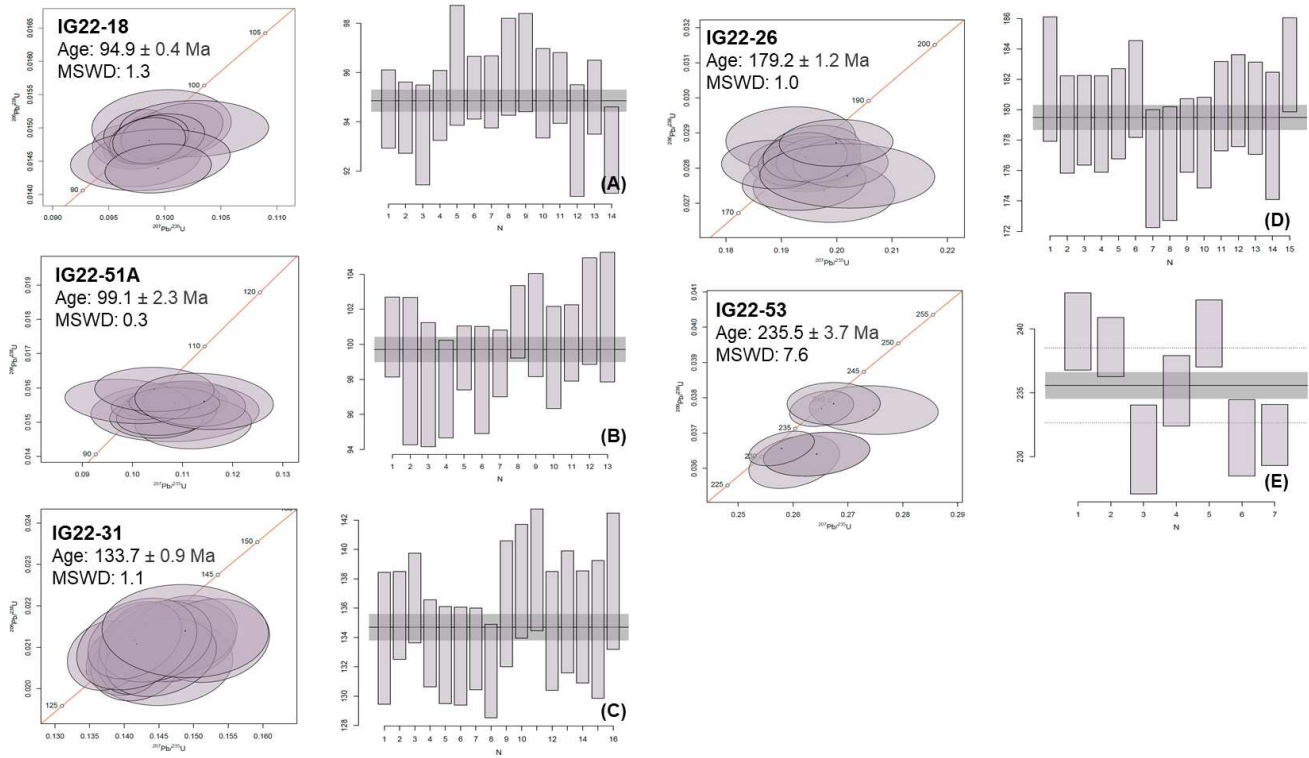


Figure 4.5 Concordia (left), weighted mean age (center), and zircon U-Pb spot (right) diagrams for selected samples, showing representative ages. Errors are expressed at 2σ . Weighted mean ages were calculated using $^{206}\text{Pb}/^{238}\text{U}$ ratios. Plots were generated using ISOPLOT (Ludwig, 2003).

Mesozoic ages obtained in the present study are distributed as follows: i) one Triassic sample (235.5 ± 3.7 Ma) from the ATB; ii) Five Jurassic samples from the eastern Ecuadorian Cordillera (Zamora Batholith) and the western Peruvian Cordillera, with ages ranging from 191.4 ± 5.3 Ma (sample IG22-23) to 157.7 ± 4.2 Ma (sample IG22-32). We report for the first time Jurassic ages in northern Peru, which we attribute to the Tauromarca and Chunchuca-Olimpo units of the western Peruvian Cordillera; iii) Four Early Cretaceous samples from the Rumpite Pluton in Peru and the Zamora

Batholith in Ecuador, with ages ranging from 139.8 ± 1.8 Ma (sample SAN18) to 126.1 ± 3.6 Ma (sample IG22-30); iv) nine Late Cretaceous samples from the Celica-Lancones arc, ranging from 100.8 ± 1.8 Ma (IG22-51B) to 87.8 ± 1.7 Ma (IG22-49).

4.3.3 Whole rock geochemistry

Fifteen samples were selected for whole-rock (wr) geochemical analysis (Table 2.4). Only one Triassic sample was analyzed in this study due to the extensive existing geochemical data available in the literature for this period (e.g., Cochrane et al., 2014; Mišković et al., 2009). In the total alkali versus silica (TAS) classification scheme (Fig. 4.6A; Middlemost, 1994), Triassic samples, including both published data and those analyzed in this study, plot within the granite, granodiorite and quartz monzonite fields. Jurassic samples are predominantly within the granodiorite field, while Early Cretaceous samples are classified as granite, granodiorite and quartz monzonite. In contrast, Late Cretaceous samples show a greater compositional diversity, ranging from granite and granodiorite to diorite and gabbro. This classification highlights that the majority of samples analyzed are felsic in composition, with only a few showing intermediate to mafic characteristics.

In general, felsic rocks display composition ranges of 63.4–75.1 wt% SiO₂, 13.5–17.1 wt% Al₂O₃, 0.24–3.9 wt% MgO, and 0.62–7.98 wt% CaO, while intermediate to mafic rocks range from 49.1–59.9 wt% SiO₂, 18.1–19.5 wt% Al₂O₃, 1.7–7.8 wt% MgO, and 7.2–14.1 wt% CaO (all values recalculated as anhydrous).

The La/Yb and Sr/Y ratios, which are commonly used as proxies for crustal thickness (Chapman et al., 2015; Profeta et al., 2015) show a uniform trend from Triassic to Cretaceous (Figs. 4.7A and 4.7B).

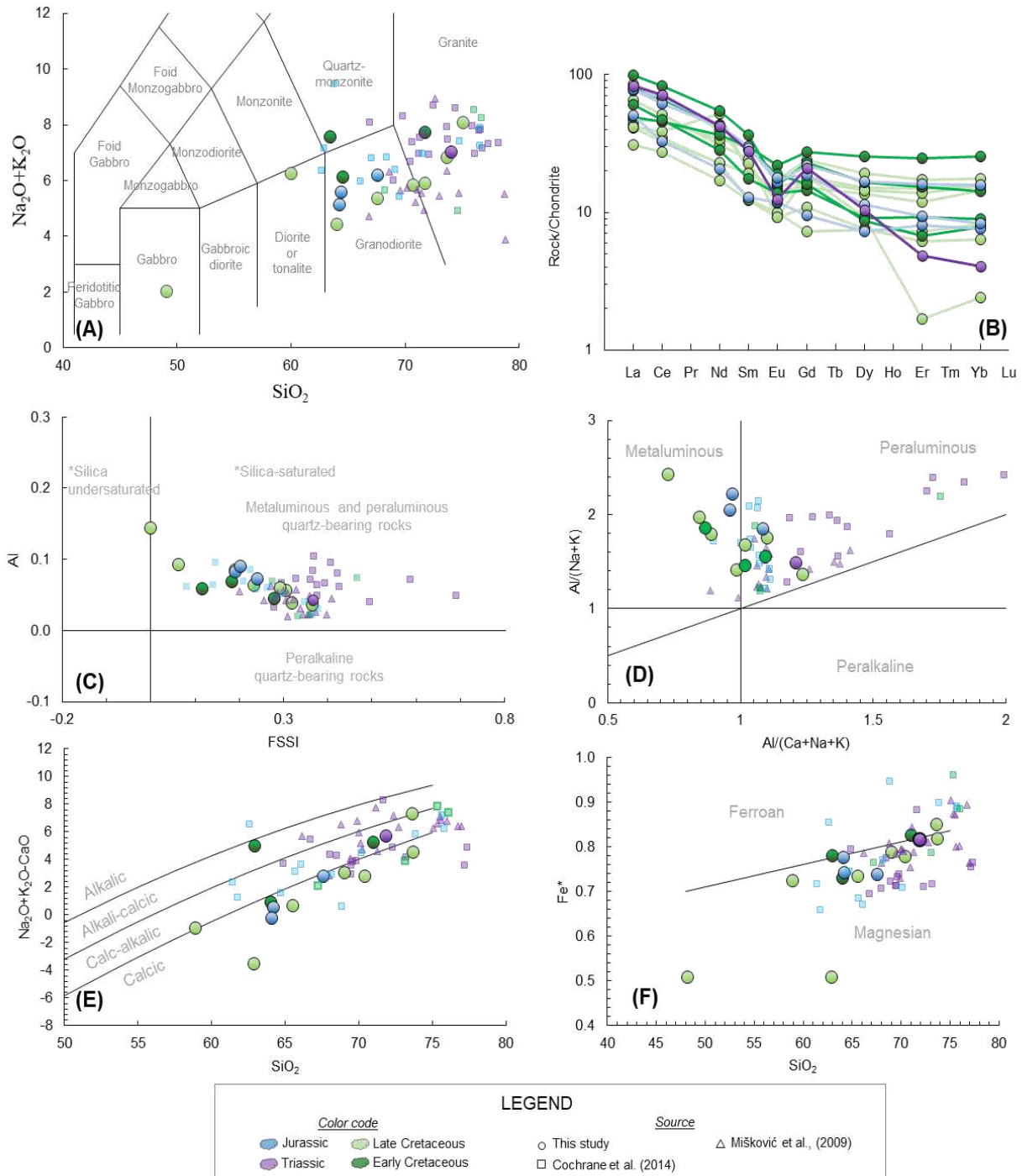


Figure 4.6 Whole rock classification diagrams. (A) Total alkali versus silica plot (Middlemost, 1994). (B) Chondrite-normalized rare element plot, normalizing values are from Sun and McDonough (1989). (C) Plot of alkalinity index (AI) vs feldspathoid silica-saturation index (FSSI). (D) Molar A/CNK versus molar A/NK plot. (E) Na₂O+K₂O-CaO as a function of SiO₂, fields are from Frost & Frost, 2008. (F) Fe* versus SiO₂ plot (Frost et al. 2001).

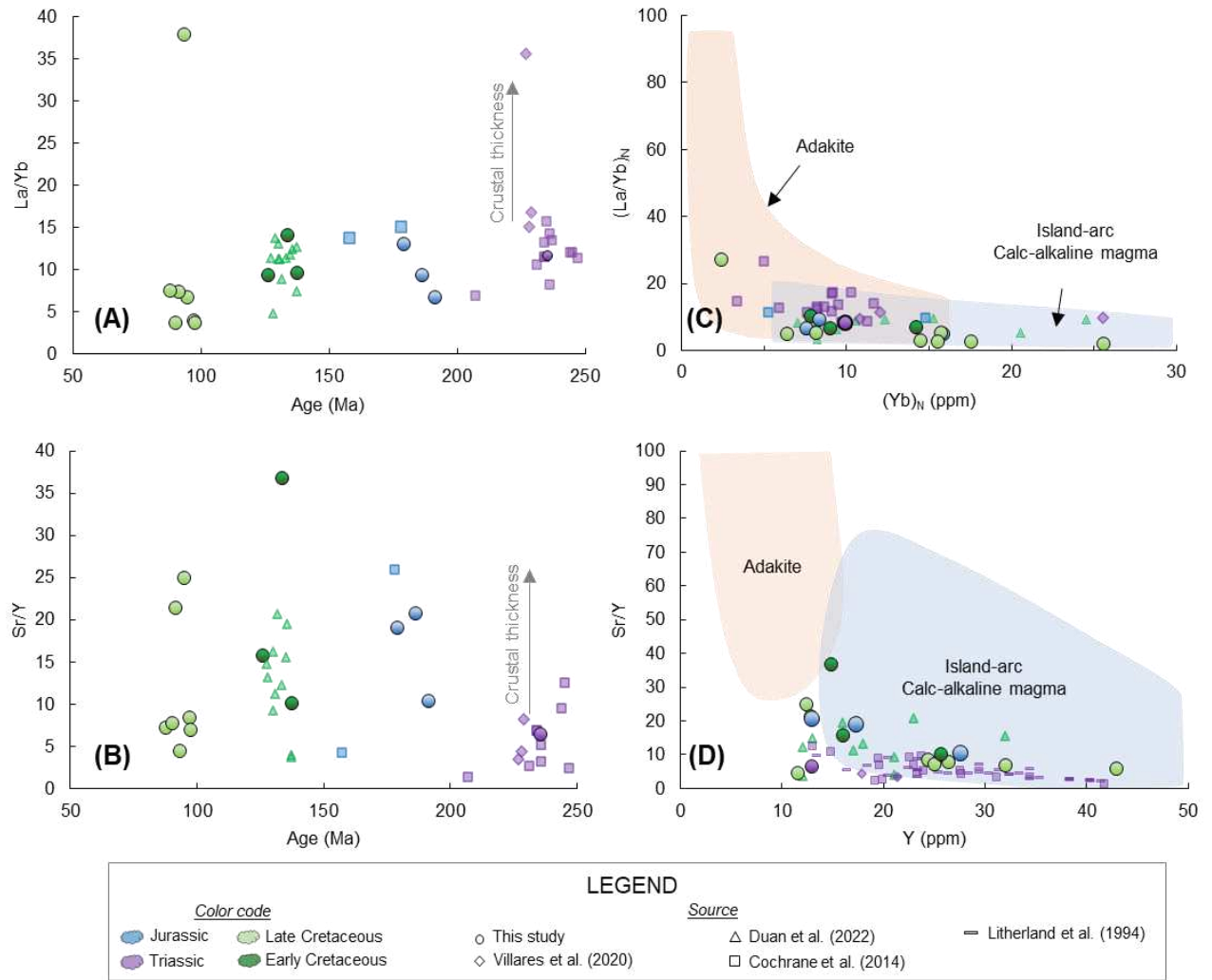


Figure 4.7 Whole-rock petrogenetic proxies related to crustal thickness and adakite discrimination diagrams. (A) La/Yb vs. Age. (B) Sr/Y vs. Age. (C) (La/Yb)_N vs. Yb_N (after Martin, 1986; N = chondrite-normalized). (D) Sr/Y vs. Y (after Drummond & Defant, 1990).

4.3.4 Zircon isotopic data

A total of 130 new $\epsilon\text{Hf}_{(t)}$ analyses are presented in Table 2.5, with the results displayed in Fig. 4.8A. $\epsilon\text{Hf}_{(t)}$ values from the Triassic samples range from -12.8 to +1.2. Jurassic samples show a range of +4.5 to +7.5, with an outlier at -17.8. Early Cretaceous values vary from +4.4 to +11.8, while samples from late Cretaceous show the highest values, fluctuating between +8.5 and +14.8, with three outliers at -10.1, -1.9 and +3.1. In general, $\epsilon\text{Hf}_{(t)}$ values show a depleting trend over time, from about 250 to about 85 Ma. However, in the Late Cretaceous (~100 Ma) they reach their maximum values, approaching depleted mantle signatures, before showing a slight decrease towards ca. 85 Ma, although the trend is subtle.

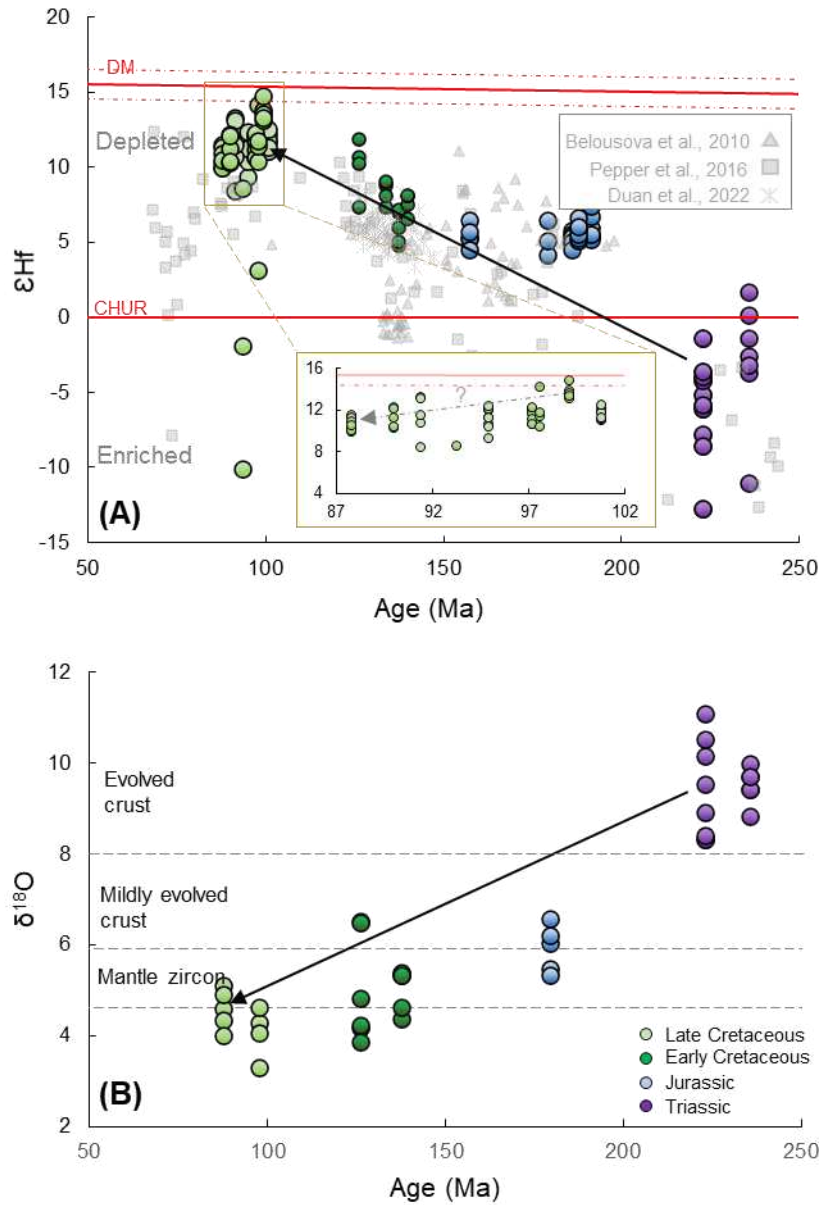


Figure 4.8 Isotopic data for intrusive samples from the of south Ecuador and north Peru. (A) ϵ_{Hf} vs. Age, including published ϵ_{Hf} isotopic data from detrital zircons by Belousova et al. (2010) and Pepper et al. (2016). (B) $\delta^{18}\text{O}$ vs. Age. Limits are from Valley et al., (2005).

A total of 44 new $\delta^{18}\text{O}$ measurements are reported in Table 2.6, and the results are displayed in Fig. 4.8B. The $\delta^{18}\text{O}$ values for the Triassic show supracrustal signatures ranging from 8.3 to 11 ‰, with an outlier at 4.7 ‰ (This study; Sandoval-Espinel et al., 2025). Samples from the Jurassic exhibit relatively consistent mantle-like values, ranging from 5.3 to 6.2‰, with an outlier of 9.5‰. Throughout the early Cretaceous, the $\delta^{18}\text{O}$ values range from 3.9 to 6.5‰. Finally, the late Cretaceous samples show values ranging from 3.3 to 5, with two values exceeding 7‰. The $\delta^{18}\text{O}$ data reveal a clear decreasing trend over time, suggesting an increasing influence of mantle-like signatures.

4.3.5 Zircon geochemistry

The zircon geochemical dataset comprises 121 measurements (Table 2.7), which exhibit considerable variability in several elements, including Dy (25-864 ppm), Gd (3-231 ppm), U (42-8130 ppm) and Y (285-9140 ppm). Chondrite-normalized trace elements patterns (Fig. 4.9A) display the typical trend for continental arc samples (Hoskin & Schaltegger, 2003) with significant enrichment in HREE to MREE relative to LREE ($(Yb/Sm)_{CN} = 44-118$; $(Yb/Gd)_{CN} = 13-39$) and negative Eu anomalies ($Eu/Eu^* = 0.20-0.31$).

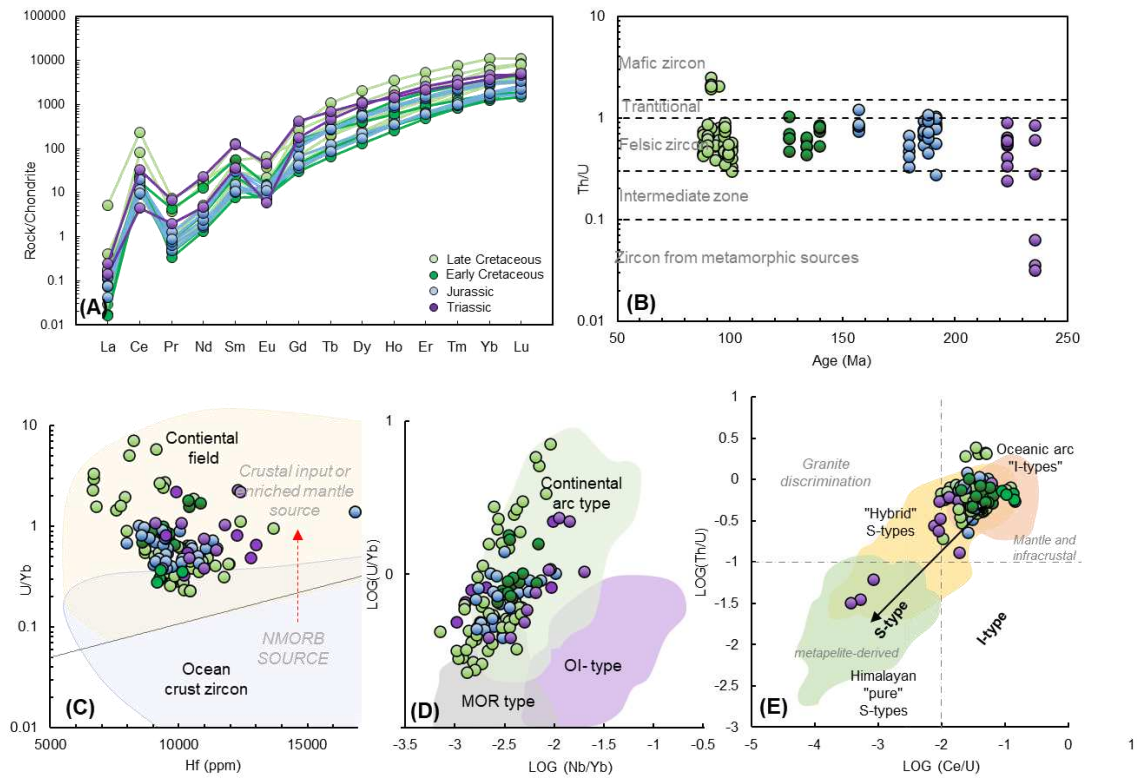


Figure 4.9 Zircon-based classification diagrams. (A) Chondrite-normalized rare element plot, normalizing values are from Sun and McDonough (1989). (B) Th/U versus age plot. (C) U/Yb vs Hf, fields for continental and oceanic zircon are from Grimes et al., (2007) (D). Tectonic discrimination diagram, boundaries are from Grimes et al., (2015). (E) Th/U vs Ce/U plot defining I-type, S-type and hybrid S-type granitoids, boundaries are from Roberts et al. (2024). MOR: Mid ocean ridge. OI: ocean island.

Zircons exhibit Th/U ratios ranging from 0.03 to 2.5 (Fig. 4.9B). High Th/U values are indicative of a magmatic origin. However, some Triassic crystals display lower Th/U values (0.03- 0.06), which are indicative of metamorphic environments (Th/U<0.1; Rubatto, 2002). Furthermore, the majority of zircons are derived from felsic rocks (Th/U>0.1; Grimes et al., 2015; Teipel et al., 2004), with the exception of some early Cretaceous crystals, for which high Th/U values (Th/U = 1.9 – 2.5) suggest a mafic source. Additionally, most of the zircons show characteristics consistent with formation within a continental arc. Although some Late Cretaceous zircons plot within the mid-ocean ridge field

(Fig. 4.9C and Fig. 4.9D), the overlap of contour boundaries in the diagram suggests that the majority was likely formed in a continental setting.

The Triassic zircons display S-type to "hybrid" S-type signatures, according to the classification schemes proposed by Roberts et al. (2024), consistent with the findings of Aspen et al. (1992) and (Riel et al., 2014), while Jurassic to Late Cretaceous zircons show signatures typical of I-type granitoids (Fig. 4.9E).

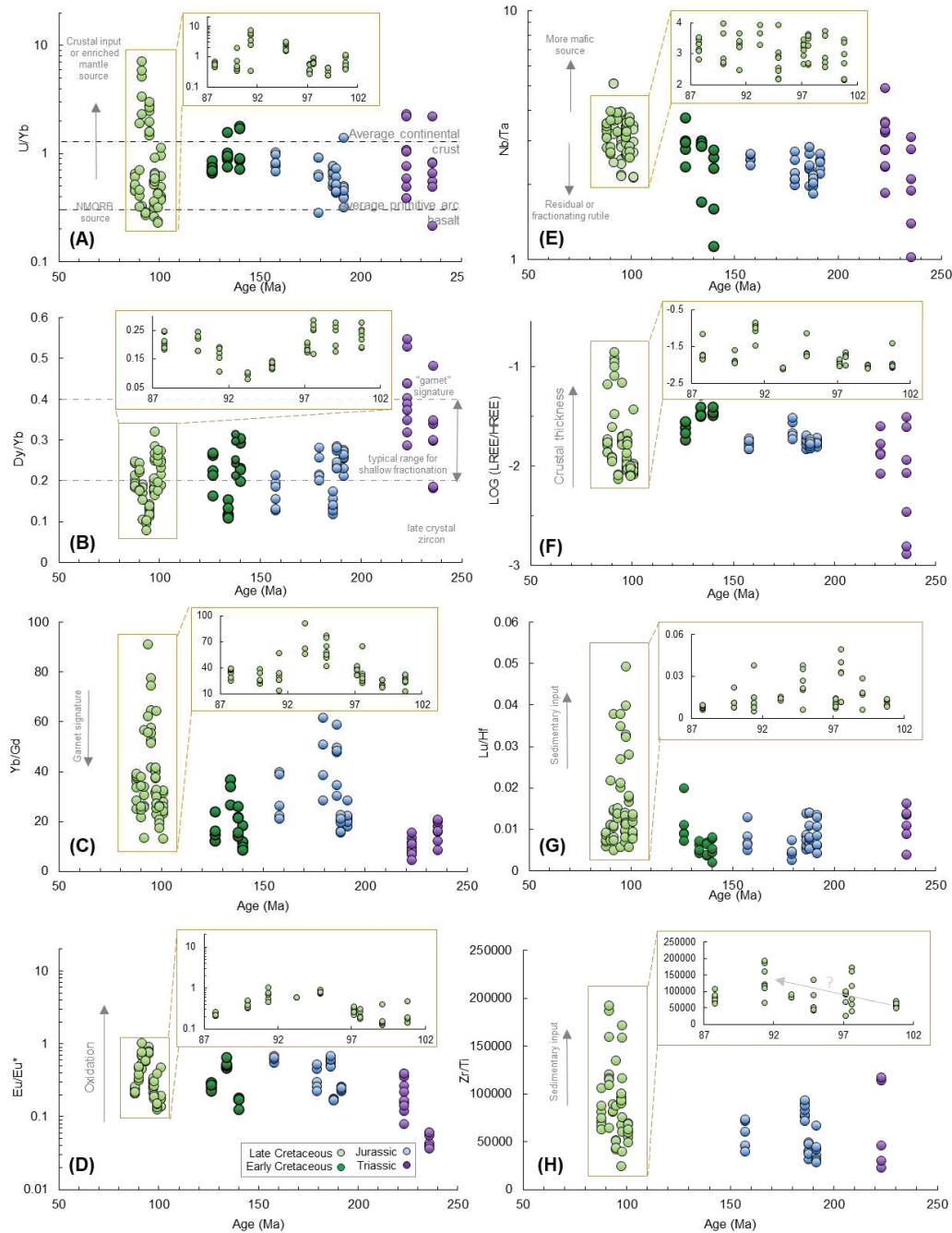


Figure 4.10 Zircon bivariate plots showing various proxies. (A) Age vs. U/Yb as a proxy for melt maturity, with domains from Grimes et al. (2015). (B-F) plots of proxies for crustal thickness, including (B) Dy/Yb vs. Age, (C) Yb/Gd vs. Age, (D) Eu/Eu* vs. Age. (E) Nb/Ta vs. Age, and (F) LREE/HREE vs. Age. (G-H) plots of proxies for sediment input, including (G) Lu/Hf vs. Age, and (H) Zr/Ti vs. Age.

The U/Yb ratio is used to distinguish between continental and oceanic zircons and assessing crustal contributions in magmas, as it reflects depletion or enrichment and the extent of crustal assimilation (Grimes et al., 2007; Grimes et al., 2015). The U/Yb ratio (Fig. 4.10A) shows a progressive increase from the Triassic to the early Cretaceous, indicating gradual enrichment. However, this trend is disrupted in the late Cretaceous, where zircons exhibit both high and low U/Yb values, suggesting derivation from a mixed source. Notably, the high Th/U values observed are primarily associated with sample IG22-19, which also presents anomalously high U concentrations. This sample is the most inland of the Late Cretaceous group.

Ratios such as La/Yb, Yb/Gd, Nb/Ta and LREE/HREE are considered as indicators of melting depth, as they reflect pressure-dependent mineral fractionation in key phases (e.g., garnet, amphibole, clinopyroxene and plagioclase) serving as proxies for crustal thickness (Drabon et al., 2024; Grimes et al., 2015; Linnen et al., 2014; Sundell et al., 2022; Trail et al., 2024). Analyzed zircons exhibit a progressive increase in Nb/Ta and LREE/HREE from the Triassic to the Early-Middle Cretaceous, though the trend is less clear in the Late Cretaceous. Additionally, Dy/Yb decreases over the same interval, while Yb/Gd shows no distinct trend (Fig. 4.10).

4.4 Discussion

4.4.1 Triassic crustal reworking

Triassic magmatism is recorded between 247.2 ± 4.3 Ma and 203 ± 1.4 Ma, and is associated with the Moromoro, Sabanilla, and Tres Lagunas units along the western flank of the Ecuadorian Eastern Cordillera, as well as the Marcabelí, La Victoria, and La Florida granitoids of the ATB (e.g., Cochrane et al., 2014; Feininger & Silberman, 1982; Litherland, 1994; Noble et al., 1997; Sandoval-Espinel et al., 2025). Further east, Triassic magmatism has also been identified in the Chinchiná unit, reaching as far as $\sim 75.5^\circ\text{W}$ (e.g., Cochrane et al., 2014; Litherland, 1994; Table S1), suggesting the existence of a wide magmatic arc. Our U-Pb zircon data support and refine these observations, documenting crystallization ages ranging from 235.5 ± 3.7 Ma to 222.8 ± 1.2 Ma, in agreement with previous reports for the ATB (Figs. 4.2 and 4.11).

Triassic magmatism (247–203 Ma) along the western margin of South America is commonly associated with lithospheric thinning and extension, although its driving mechanisms remain debated. Two main models have been proposed: (1) continental rifting linked to the breakup of Pangaea, where magmatic underplating and crustal thinning eventually led to the formation of oceanic lithosphere between northwestern South America and the Central American basement (e.g., Spikings et al.,

2015); and (2) changes in the subduction dynamics, including slab steepening and rollback, which facilitated continental crust thinning and extension (Riel et al., 2018; Villares et al., 2020).

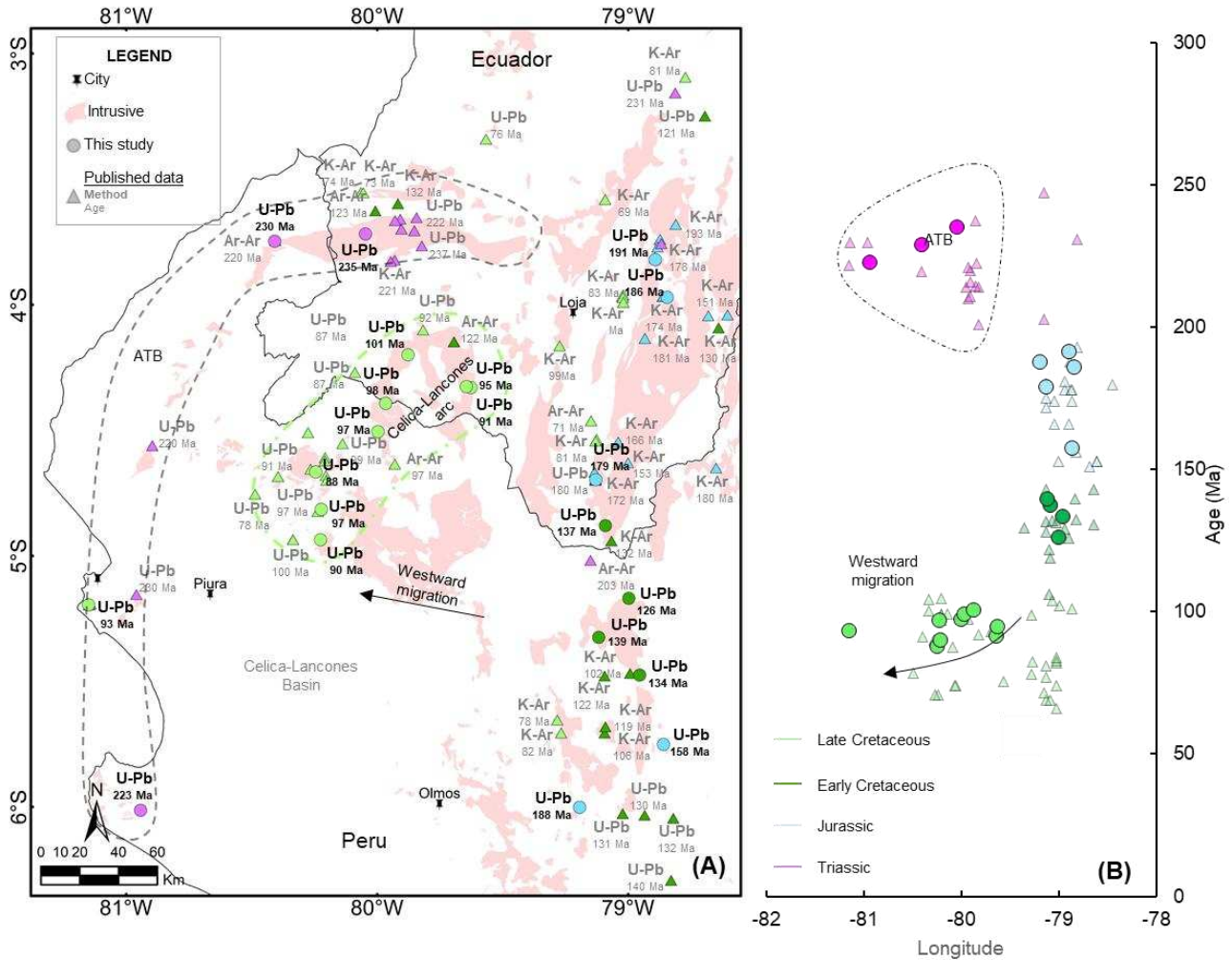


Figure 4.11 (A) Sketch of the area showing chronological constraints. (B) Ages plotted against longitude.

Geochemical and isotopic evidence from our dataset provides insights into these processes. The Triassic granitoids analyzed in this study exhibit strong crustal assimilation, as indicated by high $\delta^{18}\text{O}$ (up to 11‰) and low ϵHf_t (-12 to +2) values (Fig. 4.8), suggesting prolonged crustal reworking that persisted until at least ~220 Ma. These isotopic characteristics are consistent with a well-documented phase of anatexis during this period (Aspen et al., 1992; Villares et al., 2020), likely driven by the partial melting of low- to medium-grade semipelitic metamorphic rocks (Aspen et al., 1992).

The combination of low CaO and Na₂O contents, high Al₂O₃ concentrations, elevated $\delta^{18}\text{O}$ values, low Ce/U and Th/U ratios in zircon, and negative ϵHf_t values collectively suggest an S-type affinity for the Triassic granitoids (235–222 Ma; Fig. 4.9E). This observation indicates a substantial contribution from supracrustal (metasedimentary) sources and provides further evidence for extensive

crustal reworking. This interpretation is consistent with the models of Riel et al. (2018) and Villares et al. (2020), which document S- and A-type magmatism along the western Gondwana margin between the same interval. Conversely it differs from the model of Spikings et al. (2015), who argue that there is no evidence for significant assimilation of continental crust after ~225 Ma.

Furthermore, our data indicate that Triassic granitoids exhibit calc-alkalic geochemical signatures (Fig. 4.6F), characteristic of arc environments (Frost & Frost, 2023; and references therein). High Dy/Yb, and low Yb/Gd ratios (Fig. 4.10E) in zircons indicate amphibole stability in the source residue, consistent with deep crustal melting in a moderately thickened crust of about 30 km (Drabon et al., 2024; and references therein). Some Triassic samples also exhibit high whole -rock La/Yb ratios (Figs. 4.7A and 4.7C), suggesting a thick crust, with some compositions approaching adakite-like signatures. However, these features are not fully diagnostic of adakitic magmatism, as Sr/Y ratios do not consistently reflect the over-thickening of the crust typically associated with true adakites (Figs. 4.7B and 4.7D). Therefore, we propose that Triassic rocks formed in an intra-arc extensional setting, where partial melting of pre-thickened crust, likely driven by mantle upwelling, facilitated crustal anatexis as suggested by Aspen et al. (1992), generating the observed magmatic signatures (Fig. 4.12A).

Following this event, the region experienced a period of magmatic and tectonic quiescence, marked by the development of extensive carbonate platforms, persisting until the Early Jurassic, when subduction activity resumed (Jaillard et al., 2000; Jaillard & Soler, 1996).

4.4.2 Jurassic magmatism: from strong crustal signatures to predominantly mantle inputs

The resumption of magmatic activity during the Jurassic is evidenced in northwestern South America by the presence of the Norosí, San Martín, Segovia, Ibagué, and Mocoa-Garzón batholiths in Colombia, and the Rosa Florida, Abitagua, Azafrán, Chingual and Zamora batholiths in Ecuador (Bayona et al., 2020; Bustamante et al., 2016; Cochrane et al., 2014; Leal-Mejía et al., 2019; Spikings et al., 2015).

In southern Ecuador, Jurassic magmatism is primarily represented by the Zamora Batholith, with ages (cooling and crystallization) constrained by multiple geochronological techniques, including K-Ar in biotite and hornblende (201 - 151 Ma; Gendall et al., 2000; Litherland, 1994), U-Pb in zircon (180 - 145 Ma; Chiaradia et al., 2009; Cochrane et al., 2014; Drobe et al., 2013), Re-Os (~155 Ma- 145 Ma; Chiaradia et al., 2009; Drobe et al., 2013), and ^{40}Ar – ^{39}Ar dating on hornblendes (160–150 Ma; Chiaradia et al., 2009; Coder, 2003). Our new U-Pb zircon ages, ranging from 191 ± 5 Ma to 157 ± 4.2 Ma, fall within this interval. Furthermore, our data provide the first direct evidence of Jurassic

magmatism in northern Peru, with samples IG22-32 and IG22-34 yielding ages of 157.7 ± 4.2 Ma and 187.8 ± 3.2 Ma respectively. These results confirm that the Jurassic magmatic arc extended further south into northern Peru, indicating that Jurassic magmatic activity was more widespread than previously recognized, reaching at least as far south as $\sim 6^\circ\text{S}$.

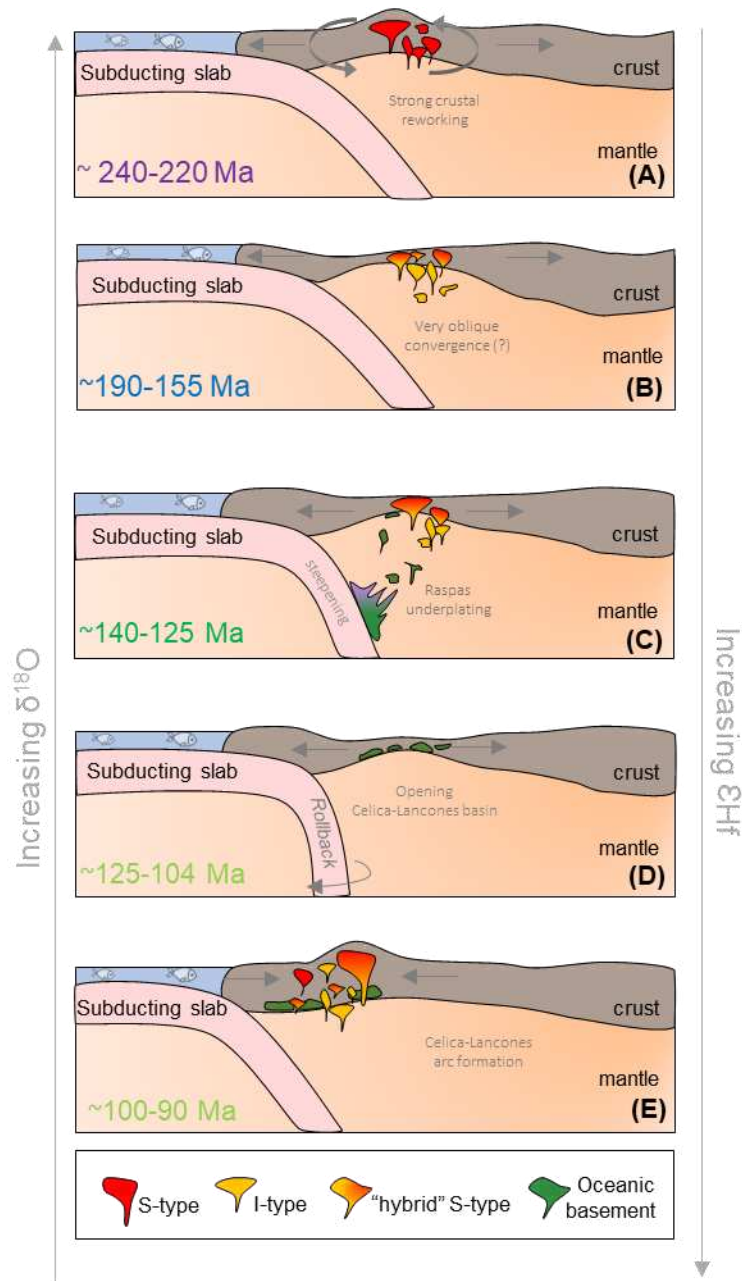


Figure 4.12 Sketch of the Mesozoic tectonic framework of southern Ecuador and northern Peru. The magmatic arc remained stable and active at ~ 190 Ma and ~ 125 Ma, before migrating westward by ~ 110 Ma, likely due to slab rollback associated with the opening of the Celica-Lancones Basin.

Additionally, samples IG22-22, IG22-23, IG22-26, and IG22-34 from the Ecuadorian Zamora Batholith, with Early Jurassic ages ranging from ca. 191 to 179 Ma, temporally overlap with the Norosí and San Martín batholiths (189–182 Ma) and the Mocoa-Garzón batholith (180–172 Ma) in

Colombia (e.g., Leal-Mejía et al., 2019; and references therein). These results challenge the previously proposed southward younging of magmatism (i.e., Spikings et al., 2015), which was based on a limited dataset from southern Ecuador and Northern Peru.

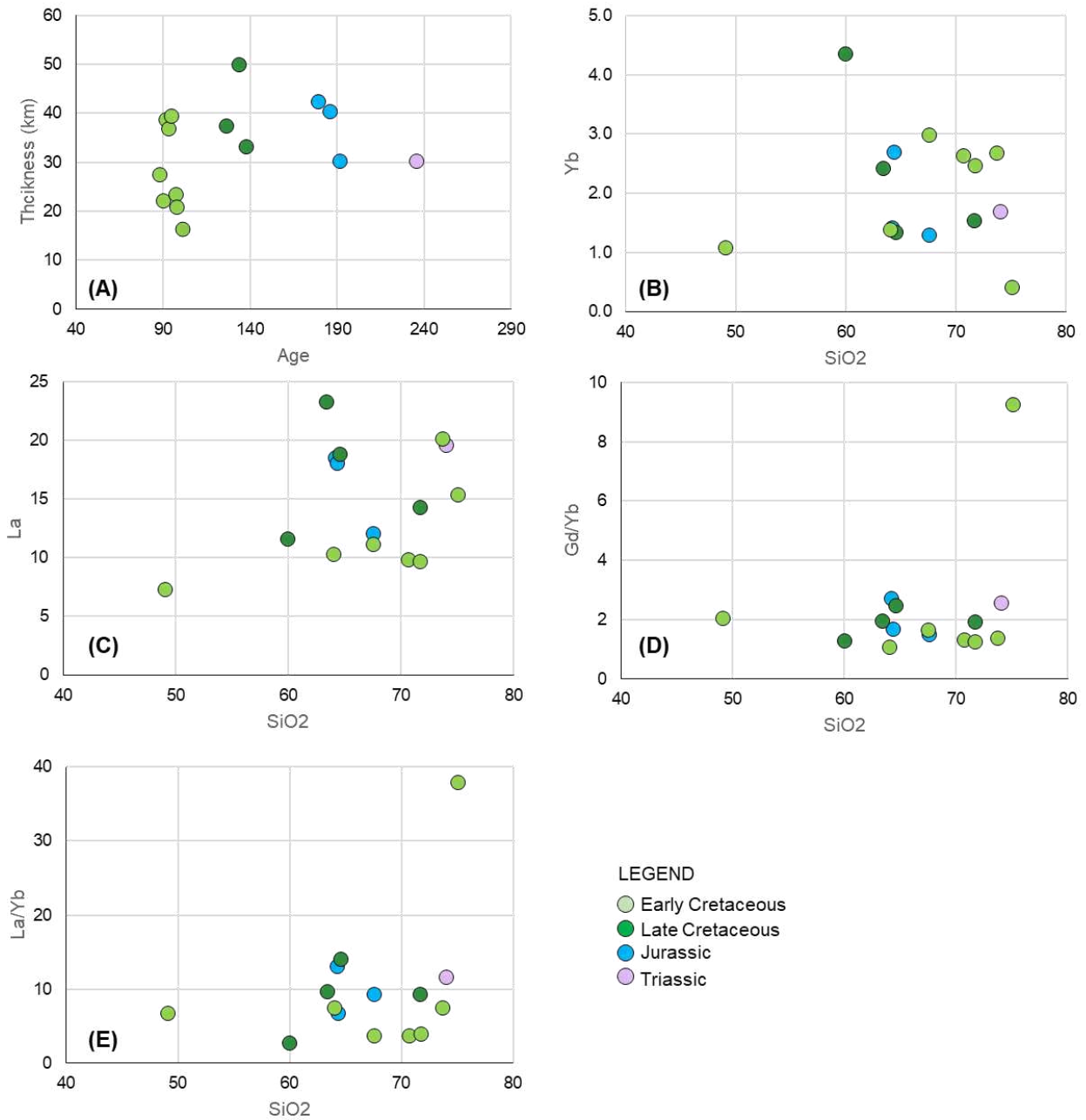


Figure 4.13 Whole-rock bivariate plots. (A) Crustal thickness versus sample age, with thickness estimates based on Sundal et al. (2021). (B) Yb (ppm) versus SiO₂ (wt%, recalculated to anhydrous basis). (C) La (ppm) versus SiO₂ (wt%, recalculated to anhydrous basis). (D) Gd/Yb versus SiO₂ (wt%, recalculated to anhydrous basis). (E) La/Yb versus SiO₂ (wt%, recalculated to anhydrous basis).

Our Jurassic isotope data reveal intermediate $\epsilon_{\text{Hf}(t)}$ values (4.2 - 7.5; Fig. 4.8A) and a shift in $\delta^{18}\text{O}$ values from high (>8‰) to 5.3–6.5‰ (Fig. 8B), indicating a transition toward more primitive compositions and increasing mantle contribution over time. Jurassic magmatism is characterized by

magnesian and calcic geochemical signatures (Figs. 4.6E, 4.6F) along with a compositional shift from peraluminous to metaluminous (Figs. 4.6C, 4.6D), in agreement with the previous suggestion of an increasing involvement of depleted magma sources (e.g., Frost & Frost, 2008). This geochemical evolution is further reflected in a shift from S-type to I-type granitoids (Fig. 4.9E), accompanied by decreasing Dy/Yb ratios (Figs. 4.10B, 4.10F), indicating melting in a garnet-free, amphibole-stable source at mid-crustal depths (Fig. 4.4C; ~35 km) (Drabon et al., 2024; Grimes et al., 2015; Tang et al., 2020). These crustal thickness estimates, based on Sundell et al. (2021), indicate values of approximately 35 km during the Jurassic, similar to those reported for the Triassic. This overall uniform trend is supported by the lack of significant variation in La, Yb, La/Yb, and Gd/Yb ratios relative to SiO₂ content (Fig. 4.13). In addition, most rocks lack the adakite-like signatures typically associated with over-thickened crusts (Figs. 4.7C and 4.7D). This trend appears to persist into the Early Cretaceous, before decreasing to around 30 km in the Late Cretaceous, as discussed later (Fig. 4.13).

Based on our geochemical dataset, we propose a broader pattern of Jurassic intra-arc extension, in which continued lithospheric thinning facilitated mantle-derived magmatism while limiting crustal recycling (Fig. 4.12B). A Jurassic extensional regime has already been documented in the northern Andes (e.g., Bustamante et al., 2016; Cochrane et al., 2014; Spikings et al., 2015). For instance, Spikings et al. (2015) and Cochrane et al. (2014) investigated Jurassic magmatism in the Ecuadorian Andes and suggested that, after ~194 Ma, slab rollback induced westward arc migration, from the Cordillera Real to the Western Cordillera, leading to the emplacement and development of a long-lived magmatic arc that persisted until ~140 Ma. They attributed this arc migration to crustal extension above a retreating subduction zone, which progressively reduced the degree of continental contamination of mantle-derived magmas. However, our U-Pb dataset does not support such an early timing for arc migration (see section 4.4.3).

4.4.3 Early-Middle Cretaceous dynamics: westward magmatic migration and Lancones basin evolution

Cochrane et al. (2014) and Spikings et al. (2015) reported a suite of Late Jurassic to Early Cretaceous ⁴⁰Ar/³⁹Ar ages from various mineral phases collected at western latitudes of Triassic and Early Jurassic arc remnants across northern and central Ecuador. Their dataset includes hornblende (~500–550 °C closure temperature), which likely records the timing of crystallization. However, this interpretation can vary depending on the magmatic setting and the cooling rate (i.e., slow versus rapid cooling), and thus hornblende ages may alternatively reflect cooling as well as biotite and plagioclase, which have significantly lower closure temperatures and are more susceptible to post-magmatic

thermal overprinting. Based on this composite geochronological evidence, the authors proposed a model of westward arc migration during the Late Jurassic to Early Cretaceous in the Ecuadorian Andes.

Nevertheless, this westward migration has not been clearly identified in southern Ecuador and northern Peru, where Early Cretaceous K–Ar ages ranging from ~140 Ma to ~129 Ma have been reported east of older arc remnants (e.g., CGS_INGEMMET, 2017; Hama, 1990; Litherland, 1994). These ages are primarily derived from whole-rock and biotite K–Ar analyses, both of which have relatively low closure temperatures and are prone to thermal resetting, thereby limiting their ability to robustly support or refute the arc migration model. Although a few U–Pb zircon ages (140–130 Ma; CGS-INGEMMET, 2017) have been reported, they remain scarce and are entirely absent within the key region between 4°S and 6°S. We present new U–Pb zircon ages ranging from 140 ± 2 Ma to 126 ± 4 Ma, obtained from plutonic bodies located at similar latitudes as the older arc remnants. These results indicate that, unlike proposed in previous models, the arc in this region remained spatially stable during the Early Cretaceous, with no clear evidence of westward migration.

Early Cretaceous magmatic activity coincided with a significant shift in subduction direction, from ESE to NNE around 145–140 Ma, alongside by slab steepening, which promoted an extensional regime (Fig. 4.12C; Riel et al., 2014). This tectonic setting facilitated the exhumation and underplating of the Raspas Ophiolitic Complex at crustal root depths between 130 and 120 Ma (Gabriele, 2002; Riel et al., 2014), contributing to the formation of a lower crust with oceanic affinities. The extensional regime is also reflected in the regional stratigraphy, with a widespread Early Cretaceous angular unconformity and the development of deep sedimentary basins such as the Chicama Basin in northern Peru (Jaillard & Soler, 1996) and the Oriente Basin in Ecuador (Baby et al., 2013; Barragán et al., 2005).

Early Cretaceous magmatic rocks (140–126 Ma) are characterized by zircon $\delta^{18}\text{O}$ values within the mantle assemblage (Fig. 4.8B) supporting a predominantly mantle-derived source. However, the presence of low $\delta^{18}\text{O}$ values ($<4.3\text{‰}$) indicates assimilation of hydrothermally altered crustal material, typically linked to shallow-level magmatic processes (Valley et al., 2005). This interpretation is reinforced by zircon Dy/Yb ratios (Fig. 10B), which are consistent with shallow-level fractionation or magmatic evolution influenced by amphibole fractionation (Drabon et al., 2024; and references therein). These findings align with Duan et al. (2022) who proposed that Early Cretaceous lavas in the north-central Western Cordillera were generated under upper amphibolite-facies conditions. Duan et al. (2022). also reported that widespread Early Cretaceous volcanic rocks in northwestern Peru are characterized by enrichment in LILEs and significant depletion in HFSEs.

This is consistent with our zircon trace element data, which display elevated LREE/HREE ratios and lower Yb/Gd values (Figs. 4.10E, 4.10C), suggesting fractionation within the amphibole stability field and the absence of garnet. Together, these geochemical signatures point to a progressive shallowing of magmatic reservoirs, with late-stage zircon crystallization occurring at mid-crustal depths (Davidson et al., 2007; Drabon et al., 2024).

Between 120 and 110 Ma, a brief compressional regime resulted from the increase of trench-normal convergence rates (Kennan & Pindell James, 2009; Maloney et al., 2013), driven by the opening of the South Atlantic Ocean at equatorial latitudes (~120 Ma; Eagles, 2007). No magmatic records for this period are provided in the present study, however, K-Ar dates reported by Hama (1990) suggest the occurrence of igneous activity during this time.

Intrusive bodies dated between 100 and 88 Ma are systematically located west of older intrusions dated as early as ~125 Ma (see previous section). This supports a westward migration of magmatism during the mid- to Late Cretaceous. This westward migration likely occurred between ~125 and 104 Ma, consistent with volcanic activity dated at 104 Ma by Winter et al. (2010). However, this timing contrasts with interpretations from northern-central Peru, where some studies suggest an earlier onset of magmatic migration; for instance, Duan et al. (2022) proposed that the Farallon Plate began to retreat around ~145 Ma. The ~125 Ma migration marks a shift from a previously stationary arc located at ~79°W (in present-day geographic reference), that was active between ~190 and ~125 Ma (Fig. 4.11). This migration is coeval with the opening of the NE-SW trending Celica-Lancones Basin (~110–100 Ma; Fig. 4.12D; Jaillard et al., 1999; Mourier, 1988; Winter, 2008), interpreted as the result of a rifted arc-type setting (Winter, 2008). During this process, the ATB block detached from the South American margin undergoing a ~65° clockwise rotation (Mourier, 1988; Riel et al., 2014). Therefore, based on the combination of strong extension in the overriding plate, evidenced by the opening of the Celica-Lancones Basin, and the relatively old age of the subducting slab (~60 Ma; Maloney et al., 2013), we propose that slab rollback was the primary driver of arc migration, occurring synchronously with the opening of the Celica-Lancones Basin around ~110 Ma. Although the Celica-Lancones Basin formed through back-arc-like extensional processes, it is currently located in a forearc position due to the westward migration of the arc front. This rollback-driven process is also thought to have contributed to Cretaceous intra-continental alkaline magmatism (110–82 Ma) in the Ecuadorian Oriente Basin (Barragán et al., 2005), in agreement with our interpretation.

A westward arc migration has also been documented in northern and central Ecuador and Colombia (Cochrane et al., 2014; Leal-Mejía et al., 2019; Ramos, 2009; Spikings et al., 2015; Toussaint & Restrepo, 1982). Previous studies (e.g., Cochrane et al., 2014; Spikings et al., 2015) proposed that

this shift occurred during the Jurassic-Early Cretaceous (see Section 4.4.2), based on the presence of Early Cretaceous units west of older arcs. For example, the Peltetec complex (Ar/Ar age of 134 Ma; Gabriele, 2002; Spikings et al., 2015) was identified west of Jurassic batholiths. However Villares et al. (2020) reinterpreted these Ar/Ar ages and, using U-Pb dating, proposed a Triassic age of 228.1 ± 1.7 Ma for the unit. They also suggested that the Ar/Ar ages reflect a Cretaceous thermal overprint associated with high-pressure metamorphism (~ 130 Ma; Gabriele, 2002; Riel et al., 2014), thus challenging the previously inferred Jurassic (~ 194 Ma) timing for arc migration.

Further south, Martínez-Ardila et al. (2023) identified a progressive eastward migration of the Mesozoic arc along central and southern Peru (8° - 18.4° S; Fig. 4.3C), with younger ages becoming increasingly distal from the trench from ~ 200 to 66 Ma. This trend, opposite to the westward migration observed in our study area, highlights the diverse dynamics of the Andean arc and regional variations in subduction processes during the Mesozoic.

4.4.4 Middle-Late Cretaceous evolution: the formation of a distinct arc

During the Albian (~ 100 Ma), active seafloor spreading in the South Atlantic Ocean and the westward drift of the South American Plate (Eagles, 2007) marked a shift in the geodynamic setting towards an Andean-type subduction system, transitioning from a predominantly extensional to a compressional regime (Ramos & Aleman, 2000), facilitating the establishment of a new continental arc around ~ 95 Ma (Fig. 4.12E), as supported by our U-Pb data (Fig. 4.3; Table 2.2). Our U-Pb ages (100.8–87.8 Ma) are consistent with those reported by Winter (2008) (U-Pb in zircon; 100.2 ± 0.5 to 78.6 ± 1.9 Ma), reinforcing the idea that Late Cretaceous magmatism was restricted to the Celica-Lancones arc.

Late Cretaceous magmatism spans a broad range of SiO_2 contents, with metaluminous to peraluminous affinities (Fig. 4.6D) from diorite to granite, with minor occurrences of silica-undersaturated gabbros (Fig. 4.6A). Notably, the gabbro (sample IG22-18) corresponds to one of the easternmost occurrences within the Late Cretaceous dataset and stands out due to its primitive character. Despite its radiogenic isotopic signature ($\epsilon_{\text{Hf}} > +10$), zircon analyses reveal very high U/Yb ratios, typical of crustal assimilation or fluid-enriched mantle sources. This paradox suggests that the gabbro is an enigmatic product of hydrous mantle melting in a continental arc setting, akin to tholeiitic gabbros in the Eastern Pontides Orogenic Belt (NE Turkey), where similar primitive but high-U/Yb magmas are attributed to bimodal arc magmatism (Temizel et al., 2022).

Whole-rock compositions predominantly show magnesian signatures (Fig. 4.6F), with calcic-alkalic and calcic affinities (Fig. 4.6E), marking a shift from the Early Cretaceous, during which compositions ranged from magnesian to ferroan. Dy/Yb ratios in zircon remain consistently low, as

do whole-rock La/Yb values (Fig. 4.7A), suggesting magma generation under amphibole-stable conditions within a crust of moderate thickness (~30 km). In contrast, LREE/HREE ratios exhibit substantial variability, with a few samples displaying anomalously high values. These outliers may reflect localized geochemical processes within the magma chamber, such as crystal accumulation (e.g., allanite, monazite) or late-stage enrichment due to residual melt evolution. However, around 93 Ma, Dy/Yb ratios increase, suggesting higher-pressure melting conditions possibly linked to garnet stability in the residue, while Eu/Eu* values decrease, indicating stronger plagioclase fractionation, likely under less oxidizing conditions. This combination may reflect crustal thickening and differentiation under relatively reduced conditions, although the decoupling between pressure and oxidation state warrants further investigation (Fig. 4.10F). This shift may reflect the onset of crustal thickening associated with the compressional regime. While these geochemical shifts point to a possible tectonic transition, additional data are needed to confirm whether significant crustal thickening took place during this period.

Zircon Th/U ratios (Fig. 4.9B) define two distinct groups: an intermediate group (0.3-1) associated with felsic compositions, and a high-Th/U group (>1.5) typically associated with mafic magmatism. This distinction is consistent with zircon Nb/Ta ratios (Fig. 4.10E), which tend toward values indicative of mafic sources, highlighting the significant influence of mafic components in the magma system. The broad distribution of U/Yb ratios (Fig. 4.10A) suggests contributions from both MORB-like sources and enriched mantle or crustal inputs. The coexistence of oceanic and continental zircon signatures (Figs. 4.9C and 4.9D), further supports a mixed-source model, which is also evidenced by zircon isotopic data. The highly depleted ϵHf_t values, in some cases approaching the depleted mantle (DM) field (Fig. 4.8A), along with mantle-like $\delta^{18}\text{O}$ values, indicate a significant mantle-derived component. However few low $\delta^{18}\text{O}$ values (<4.4‰), outliers within the dataset, may reflect assimilation or partial melting of gabbroic protoliths (Valley et al., 2005). Therefore, our data suggest that Late Cretaceous magmatism may have originated from: (1) melting of lower crustal material with mantle-like signatures, possibly linked to the underplated Early Cretaceous Raspas Ophiolitic Complex, or (2) melting of an upwelling mantle associated with the opening of the Celica–Lancones Basin between the ATB and the western South American margin, which favored the formation of an oceanic-like basement. At present, the spatial coverage of our dataset does not allow us to unequivocally distinguish between these two scenarios. However, given that lateral arc migration is intrinsically linked to the second scenario (i.e., the development of the Celica-Lancones basin), we favor this interpretation. Alternatively, a combination of both processes, where the formation of an oceanic-like basement during the opening of the Celica-Lancones basin coincided with the remobilization of underplated Raspas material, could also be considered. Regardless of the

mechanism, it is clear that the Celica-Lancones arc was emplaced on a basement with oceanic affinities. Therefore, the Celica-Lancones Arc, spanning the Late Albian to Turonian (100–88 Ma), could be considered as a northern counterpart to the Aptian-Albian volcanic arc of Peru due to their coeval magmatism. However, they differ in basement characteristics: the central-southern Peruvian arc was emplaced on continental crust (i.e., Cobbing & Pitcher, 1983; Martínez-Ardila et al., 2023; Soler, 1991), while the Celica-Lancones Arc formed on oceanic crust. These differences suggest that, despite their similar timing, the two arcs evolved under distinct tectonic conditions, leading to the development of different basement types. This underscores the complexity of magmatic arcs along the same subduction margin. Our findings highlight the strength of a multi-proxy approach, where the integration of U-Pb geochronology, whole-rock geochemistry, zircon isotopic analyses ($\delta^{18}\text{O}$, ϵHf), and trace element data provides a robust framework for reconstructing the evolution of magmatic systems and their tectonic context in complex orogenic settings.

4.5 Conclusions

Our U-Pb zircon data refine the geochronological framework of Mesozoic magmatism in southern Ecuador and northern Peru, providing new insights into the spatiotemporal evolution of the arc. Triassic magmatism occurred in an extensional setting and is recorded both along the Andean belt and within the ATB. Between 240 and 220 Ma, magmatism was characterized by significant crustal recycling and the formation of S-type granitoids formed by anatexis of pelitic source rocks, as indicated by our zircon Hf-O isotopic data. This suggests sustained crustal assimilation until at least 220 Ma. Following this period, post-Triassic magmatism transitioned to the emplacement of voluminous I-type granitoids. During the Jurassic (191–158 Ma), magmatism occurred under an extensional arc setting, characterized by a shift toward less radiogenic isotopic compositions, reflecting an increasing contribution from primitive mantle sources. The new Jurassic ages obtained in this study challenge the notion of a missing Jurassic arc in northern Peru, suggesting that magmatic activity may have been more widespread. Furthermore, the synchronous nature of Jurassic magmatism in southern Ecuador, northern Peru, and Colombia questions the previously established southward younging trend for the region.

In the Early Cretaceous, changes in subduction geometry, possibly associated with slab steepening, influenced magmatism, resulting in melting of mainly mantle sources. Between ~190 and ~125 Ma, the magmatic arc remained stable and active. By the middle Cretaceous (~110 Ma), the arc migrated westward, likely driven by slab rollback and coeval with the opening of the Celica-Lancones Basin, contradicting previous interpretations of an earlier Jurassic–Early Cretaceous arc migration. During the Late Cretaceous (~100 to ~88 Ma), magmatism evolved within a compressional setting and

involved contributions from oceanic-like components. The oceanic signature may have originated from either the melting of lower crustal material, possibly linked to the underplated Early Cretaceous Raspas Ophiolitic Complex, or direct input from an upwelling depleted mantle, forming an oceanic-like basement during the opening of the Celica-Lancones Basin, or a combination of both.

Declaration of Competing Interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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CHAPTER 5: CENOZOIC MAGMATIC EVOLUTION OF THE PIURA SEGMENT OF THE PERUVIAN COASTAL BATHOLITH: INSIGHTS FROM ZIRCON U–Pb GEOCHRONOLOGY, HF–O ISOTOPES, AND WHOLE-ROCK GEOCHEMISTRY

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Declaration statement

This manuscript is currently under review by the co-authors and is intended for submission to Tectonics, although the final choice of journal has not yet been confirmed.

CRedit authorship contribution statement

Sandoval Espinel, L.C.: Conceptualization, Investigation, Data curation, Methodology, Writing – original draft. **Witt, C.:** Conceptualization, Supervision, Funding, Writing – review & editing. **Iglesias, J.:** Methodology, Writing – review. **Zattin, M.:** Conceptualization, Supervision, Funding, Writing – review & editing. **Bermudez, M.:** Supervision, Funding, Writing – review & editing. **Poujol, M.:** Methodology, Data curation, Writing – review. **Bruguier O.:** Methodology, Data curation, Writing – review. **Sandoval Espinel, J-J.** Visualization, Methodology, Writing – review.

Highlights

1. First U-Pb constrains on the Paltashaco and Pamparumbe plutons yield 59–29 Ma and 44–20 Ma, respectively.
2. Isotopic heterogeneity suggests assimilation of variable basement sources.
3. Basement composition controlled magma sources; tectonic shifts modulated magmatism.

ABSTRACT

The magmatic evolution of the northern Peruvian Andes remains poorly constrained despite its key position at the Andean orocline. Here we present an integrated study combining zircon U–Pb geochronology, whole-rock geochemistry, and zircon Hf–O isotopes for Cenozoic plutons of the Piura segment within the Peruvian Coastal Batholith. Our data reveal prolonged magmatism from ~60 to 20 Ma and provide the first robust emplacement ages for major plutons, including Paltashaco (59–29 Ma), Pamparumbe (44–20 Ma) and Rumipite (40 Ma). Geochemical and isotopic heterogeneity across the region indicates assimilation of distinct basement sources. Western plutons (Mallingas, Pamparumbe, Tucto–Penachi, UNW, and western Paltashaco domain) display juvenile, mantle-like signatures ($\epsilon\text{Hf}(t)$ up to +8.2; $\delta^{18}\text{O} = 4.8\text{--}6.5\text{‰}$), reflecting derivation from a crust with oceanic affinity. In contrast, eastern plutons (Rumipite, UNE and eastern Paltashaco domain) show evolved compositions ($\epsilon\text{Hf}(t)$ down to –5.2; $\delta^{18}\text{O}$ up to 8.1‰), consistent with melting of continental crust. These patterns suggest that the pre-existing lithospheric architecture exerted the primary control on magma genesis, while Cenozoic transitions from compressional to extensional regimes modulated mantle–crust interactions.

5.1 Introduction

Igneous rocks emplaced along subduction margins are typically dominated by intermediate to felsic compositions (DeCelles et al., 2009; Murphy, 2007). These compositions commonly reflect prolonged magmatic differentiation and, in many cases, variable degrees of crustal assimilation (Ducea et al., 2015; Hildreth & Moorbath, 1988; Xu, 2023). However, arc magmatism also records the influence of several additional factors, including contributions from heterogeneous mantle sources (Iveson et al., 2021; Turner et al., 2017), variable slab-derived inputs (Walowski et al., 2016), localized tectonic processes, and inherited basement signatures (Muñoz et al., 2013). Understanding the evolution of magmatism along continental arcs therefore provides key insights into the interplay between subduction dynamics, mantle–crust interactions, and regional tectonic regimes.

The Andean margin provides an exceptional natural laboratory for investigating magmatic processes, owing to its long-lived subduction history driven by the descent of the oceanic Nazca (formerly Farallon) Plate beneath the South American Plate (Fig. 5.1; Isacks, 1988). Nevertheless, the tectonic evolution of key segments of the orogen, including northern Peru, remain poorly constrained. This region hosts a major orocline that defines the transition between the Northern and Central Andes (Ramos & Aleman, 2000), characterized by a shift in trend from N20°W to N20°E (Mourier, 1988). Northern Peru is particularly significant, not only due to its position within the orogen, but also because it hosts the intrusive bodies of the northern segment (Piura Segment) of the Peruvian Coastal

Batholith (PCB). This segment may differ from the better-constrained central and southern segments of the batholith (e.g., Arequipa and Lima; Mamani et al., 2010; Martínez-Ardila et al., 2023) in terms of magmatic evolution and geochemical characteristics, potentially reflecting distinct tectonomagmatic conditions. However, the absence of comprehensive datasets for the Piura Segment precludes robust comparison, as evidenced by the scarcity of precise radiometric ages for numerous plutons (e.g., Mallingas, Paltashaco and Pamparumbe; De la Cruz Wetzell, 1995; Jaimes Salcedo, 2011, 2012; Reyes Rivera & Caldas Vidal, 1987) and by the limited availability of geochemical and isotopic data required to characterize magmatic reservoirs and mantle–crust interactions (Kemp et al., 2006; Witt et al., 2023).

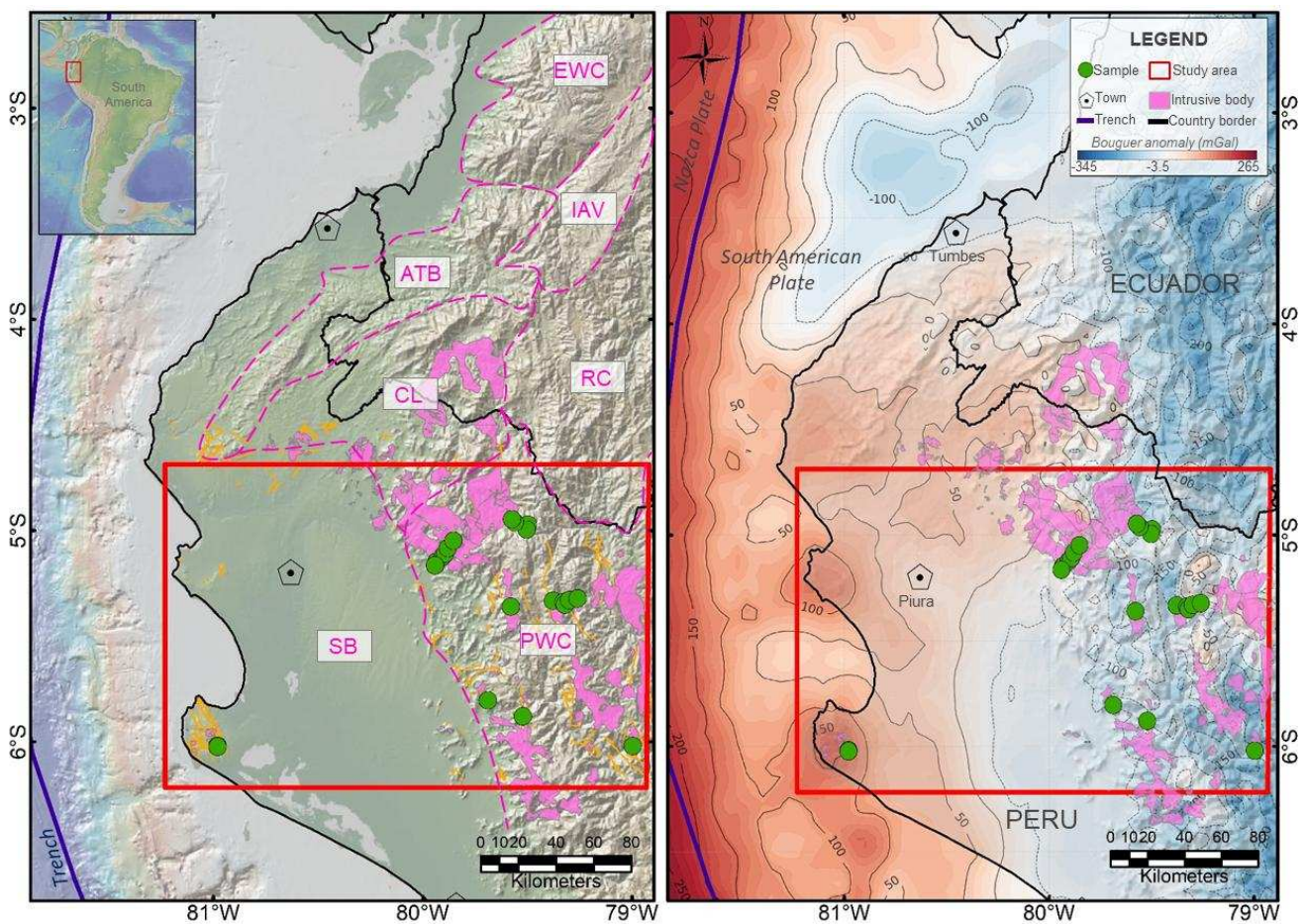


Figure 5.1 (A) Simplified geological sketch of the southernmost Northern Andes highlighting the main structural domains. (B) Bouguer anomaly map of the study area. Both panels show the intrusive bodies and sample locations analyzed in this study. Abbreviations: ATB – Amotape–Tahuin Block; CL – Celica–Lancones; EWC – Ecuadorian Western Cordillera; IAV – Inter–Andean Valley; PWC – Peruvian Western Cordillera; RC – Real Cordillera; SB – Sechura Basin.

The nature and architecture of the basement in this segment also remain controversial. Some authors argue that the Cenozoic arc was emplaced on an autochthonous continental basement dominated by Paleozoic metamorphic rocks and Triassic intrusions (Aspen et al., 1992; Witt et al., 2017). Others,

however, propose the presence of an oceanic basement formed through either arc rollback (Sandoval-Espinel et al., 2025) or underplating of ophiolitic complexes (i.e., Rasper; Riel et al., 2014).

Zircon is a ubiquitous mineral in intermediate to felsic igneous rocks, highly resistant to hydrothermal alteration and low to moderate metamorphism (Finch & Hancher, 2003). This allows zircon to effectively retain information about its parent magma (Grimes et al., 2007), making it a powerful tool for understanding the geological processes involved in the formation and evolution of its host rocks. Furthermore, advances in in-situ U-Pb dating, $\epsilon\text{Hf}_{(t)}$ and $\delta^{18}\text{O}$ analyses using LA-ICP-MS have enabled the precise determination of crystallization ages and their direct correlation with geochemical and isotopic signatures (Hawkesworth & Kemp, 2006; Witt et al., 2023).

This study presents an integrated analysis of zircon U-Pb geochronology, whole-rock and mineral geochemistry, zircon $\epsilon\text{Hf}_{(t)}$ and $\delta^{18}\text{O}$ isotopes and regional Bouguer gravity data to constrain the magmatic history of the Piura Segment of the PCB (4°S–6°S). We aim to establish precise crystallization ages for Cenozoic granitoids, characterize their magmatic sources, elucidate crust–mantle interactions and lithospheric architecture. The integration of these datasets provides new constraints on the temporal and tectonomagmatic evolution of the northern Peruvian Andes.

5.2 Geologic context

5.2.1 The Piura segment of the Peruvian Coastal Batholith

The PCB is the most extensively exposed plutonic assemblage in Peru (Cobbing & Pitcher, 1983; Pitcher, 1985). It has been subdivided mainly on the basis of their compositional and morpho-structural characteristics (Reyes Rivera & Caldas Vidal, 1987) into five petrographic segments: Toquepala (18–16.5 °S), Arequipa (16.5–12 °S), Lima (12–9 °S), Trujillo (9–6.5 °S) and Piura (6.5–4 °S; Mukasa, 1986; Pitcher, 1985).

The Piura segment is characterized by a series of plutons and minor intrusions scattered throughout the northern Peruvian Western Cordillera, defining the morphology of the region (De la Cruz Wetzell, 1995; Jaimes Salcedo, 2012; Jaimes Salcedo et al., 2013; Reyes Rivera & Caldas Vidal, 1987; Wilson, 1984). The major plutonic bodies in the area include the Upper Cretaceous–Paleocene (K–Ar = 82.5 – 64 Ma; Table S1; Hama, 1990) Diorite–Tonalite Rumipite massif; the Paltashaco complex, located near the towns of Santo Domingo and Pacaipampa (Fig. 2; Reyes Rivera & Caldas Vidal, 1987); the voluminous Pamparumbe batholith, located along the towns of Santo Domingo and Canchaque (Reyes Rivera & Caldas Vidal, 1987); the elongated Eocene (^{40}Ar – ^{39}Ar = 43 ± 2.8 ; Ulrich, 2005) Suyo pluton; the Oligocene (U–Pb = 29.8 ± 0.7 Ma; CGS_INGEMMET, 2017) Tucto–Penachi unit, which forms a roughly north–south trending body; and the Lomas, Peña Blanca, and Purgatorio

plutons, which together form a sub-circular Maastrichtian–middle Eocene body (^{40}Ar - ^{39}Ar = 71 - 46 Ma; Winter, 2008) around the town of Lomas.

5.2.2 Cenozoic tectonic framework of the Piura segment

The Cenozoic tectonic evolution of northern Peru was driven by the subduction of the Farallon plate and subsequent Nazca plate, beneath the South American continental plate (Espurt et al., 2018; Jaillard et al., 2000). From Campanian to Middle Paleocene, significant increase in the convergence rates (Maloney et al., 2013; Somoza & Ghidella, 2012) marked the onset of the compressional Andean orogeny (Espurt et al., 2018; Jaillard et al., 2000). This compressional setting persisted through the Late Paleocene and Early Eocene, during which the Andean chain underwent substantial deformation (Espurt et al., 2018; Witt et al., 2019). From middle to late Eocene, the region transitioned to an extensional regime (Espurt et al., 2018), which promoted both the widening of the magmatic arc (Jaillard et al., 2000) and subsidence of the continental forearc zones, creating a belt of marine forearcs (Salaverry and Talara; Jaillard, 2025).

During the Oligocene, the breakup of the Farallon Plate into the Cocos and Nazca Plates (Jaillard et al., 2000; and references therein), resulted in a transition to a compressional setting, leading to the formation of an eastward-trending orogenic wedge (Espurt et al., 2018; Jaillard, 2025). By the mid-Miocene or earlier, the subduction of the Inca Plateau may have induced slab flattening (e.g., George et al., 2022; and references therein), which could have contributed to the shutdown of magmatism and the gradual expansion of the volcanic arc (Skinner & Clayton, 2013; Witt et al., 2023).

This geodynamic framework has been reconstructed mainly from structural and sedimentary evidence (Espurt et al., 2018; Hungerbühler et al., 2002; Jaillard et al., 2000, 2025), while magmatic constraints remain comparatively limited (George et al., 2021; Mamani et al., 2010; Witt et al., 2017), here the available geochemical and isotopic data have generally been interpreted within the context of tectonic regime changes. However, recent evidence from southern Ecuador (Iglesias et al., in review) suggests that basement composition exerts a first-order control on magmatic signatures, potentially overriding the effects of short-term tectonic shifts.

5.3 Analytical methods

Nineteen samples were collected from intrusive bodies along the Piura segment of the PCB, from the Tucto-Penachi unit, the Paltashaco and Pamparupe plutons, and smaller unnamed intrusions. Sample locations and nomenclature are listed in Table 2.1 and illustrated in Fig. 5.2.

For mineral analysis, zircons were extracted at the Universidad Pedagógica y Tecnológica de Colombia using crushing, sieving, magnetic and heavy liquid concentration techniques. While, for whole rock analysis, samples were washed, crushed and pulverized to a particle size of $< 10 \mu\text{m}$ at the University of Lille, France.

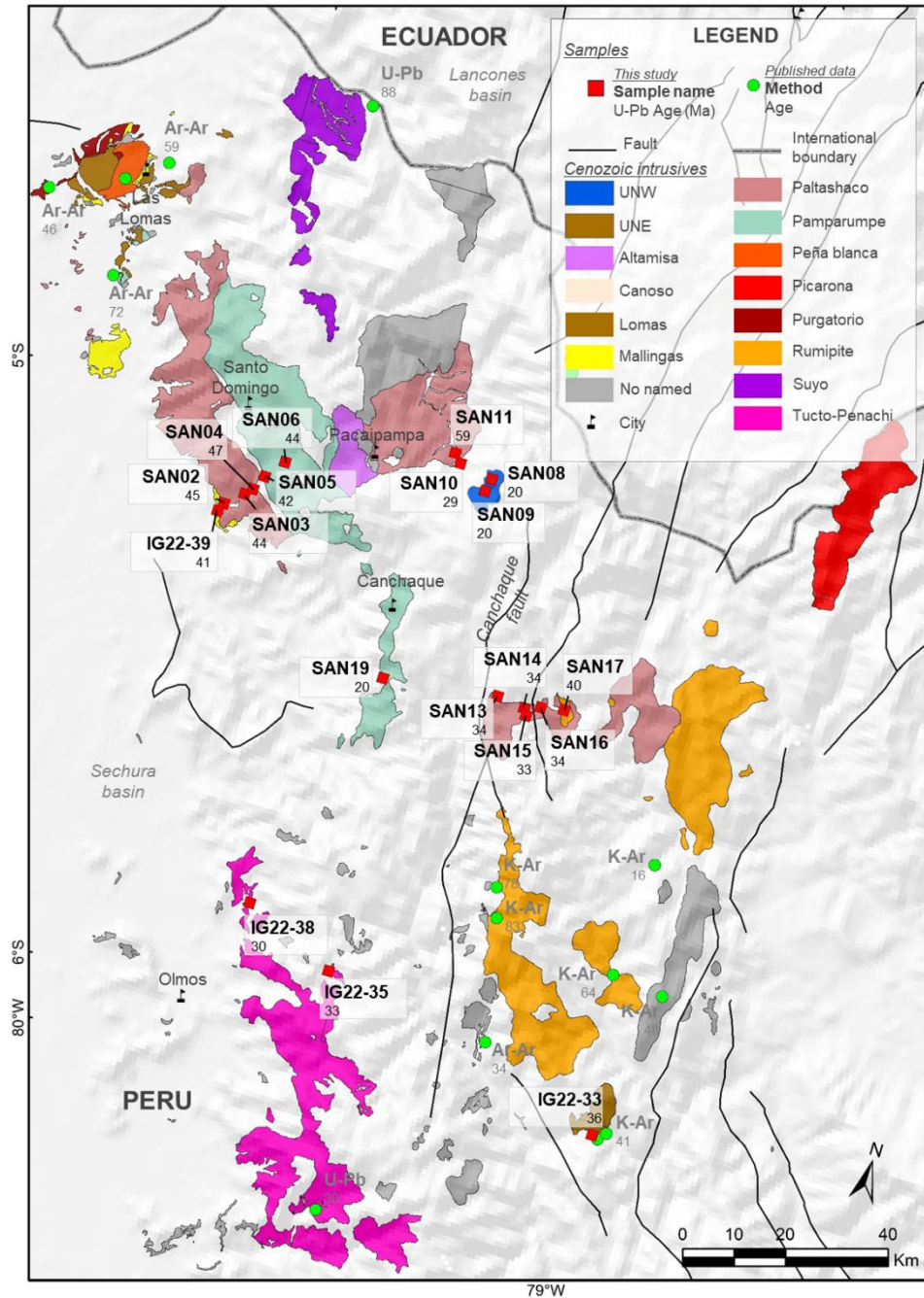


Figure 5.2 Simplified map of the Piura segment of the Peruvian Coastal Batholith, showing intrusive bodies along with published geochronological data (CGS-INGEMMET, 2017; Hama, 1990; Litherland et al., 1994; Ulrich, 2005; Winter, 2008) and sample locations from this study.

5.3.1 U-Pb geochronology

Zircon U-Pb dating was conducted using a LA-ICP-MS at the GeOHeLiS analytical platform (OSUR, Univ. Rennes) using an ESI NWR193UC Excimer laser coupled to a quadrupole Agilent 7700x ICP-MS equipped with a dual pumping system to enhance sensitivity. The instrumental setup is described in chapter 2. Data were analyzed employing the ISOPLOT package program (Ludwig, 2003). $^{206}\text{Pb}/^{238}\text{U}$ dates were used since zircons are younger than 1 Ga. Only concordant ages (90 to 110%) are presented and listed in Table 2.2. Additionally, weighted mean plots for each sample are shown in Fig. 5.3.

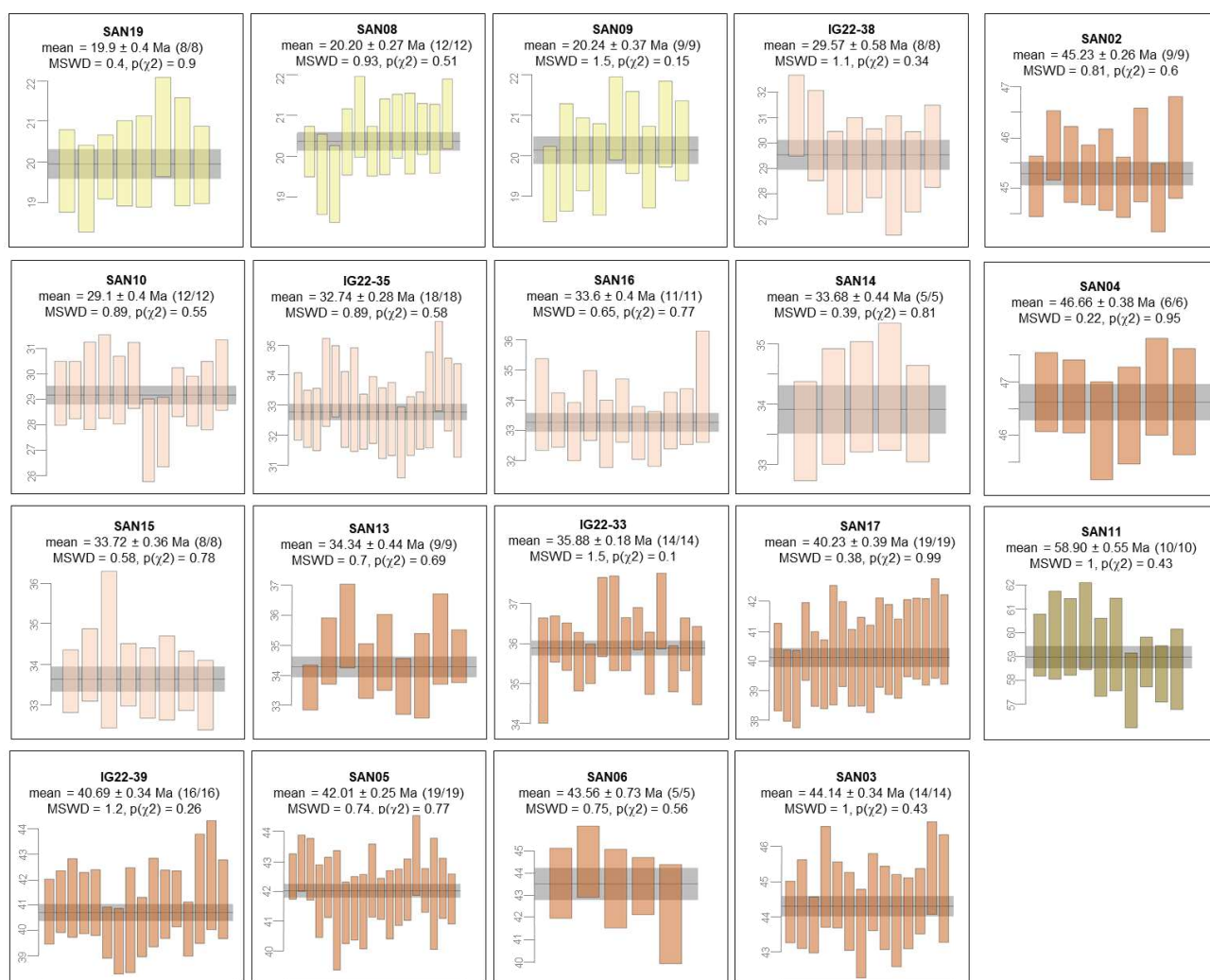


Figure 5.3 U-Pb weighted mean age plots from the analyzed intrusive rocks of the Piura segment of the PCB. MSWD: mean square of weighted deviates.

5.3.2 Petrographic analysis

Twelve thin sections were prepared at the University of Padova. Rock samples were cut, mounted on glass slides with epoxy resin, and ground to $\sim 30\ \mu\text{m}$ thickness using progressively finer abrasives; thickness was verified by quartz interference colors under polarized light. Petrographic analyses were

conducted under plane- and cross-polarized light at 5×, 10×, and 20× magnifications to document mineral assemblages, textures, and grain morphologies. Modal mineralogy was quantified using a point-counting method along an S-shaped traverse, recording 200 points sample. Representative photomicrographs in Figure 5.5.

5.3.3 Whole rock geochemistry

Whole-rock major and trace elements were determined for twelve samples at the Institut Universitaire Européen de la Mer, Université de Bretagne Occidentale (Brest, France) using a Jobin Yvon Ultima 2 inductively coupled plasma atomic emission spectrometer (ICP-AES). Sample preparation followed the procedure outlined by Cotten et al. (1995), which included digestion with HF and HNO₃, neutralization with boric acid, and subsequent dilution in nitric acid. Calibration was performed using international standards (AC-E, BEN, JB-2, PM-S, and WS-E) along with internal standards (MORB-E, CB2, CB15, and CB18), ensuring reliable external reproducibility. Relative standard deviations are < 2%, except for low values (< 0.50 wt.%), for which the absolute standard deviation is ± 0.01 wt.%. The results are presented in Table 2.4.

5.3.4 Trace elements and Hafnium analyses in Zircon

Zircon trace-element concentrations and Hf isotopes were measured simultaneously by laser ablation split-stream analyses (LASS-ICP-MS). Measurements were achieved at Géosciences Montpellier (University of Montpellier, AETE-ISO regional facility of the OSU OREME); instrumentation and analytical procedures are detailed in chapter 2. Trace element data reduction was carried out using the GLITTER software package (van Achterberg et al., 2001), with results summarized in Table 2.7. Hafnium isotopic compositions are reported in epsilon (ε) notation, expressing deviations of measured ¹⁷⁶Hf/¹⁷⁷Hf ratios from the Chondritic Uniform (CHUR; Vervoort et al., 1999), and are presented in Table 2.5.

5.3.5 δ¹⁸O analyses

Zircon oxygen isotopic compositions were analyzed using a Cameca IMS 1270 E7 ion microprobe at the Centre de Recherches Pétrographiques et Géochimiques, Université de Lorraine, France. The analyses used a Cs⁺ primary ion beam accelerated at 10 kV with an intensity of 2.5 nA. Instrumental mass fractionation values ranged from 4.41‰ to 6.418‰, determined by BrF₅ extraction and gas source mass spectrometry. Further technical details can be found in Witt et al. (2023). δ¹⁸O data are reported in Table 2.6.

5.3.6 Bouguer gravity anomalies

The Bouguer gravity anomaly data were derived from the BGI EGM2008 global model. In this model, Bouguer and topographic corrections were applied using the FA2BOUG algorithm (Fullea et al., 2008), with terrain corrections calculated to a radius of 167 km using the ETOPO1 global DEM (1 arc-minute resolution) and a standard reduction density of 2.67 g cm^{-3} . The discrete data points were interpolated onto a regular grid using Python to generate a continuous gravity anomaly surface. This grid was contoured at 50 mGal intervals to highlight regional gravity patterns. The final georeferenced raster was imported into ArcGIS 10.5 and combined with the sample locations for interpretation (Fig. 5.1B).

5.4 Results

5.4.1 U-Pb geochronology

Nineteen radiometric ages were obtained from plutonic intrusions along the Piura segment of the PCB. Analyzed zircons from all samples were predominantly euhedral to subhedral, and show one or both terminations (Fig. 5.4). Weighted mean plots for representative samples are shown in Figure 5.3.

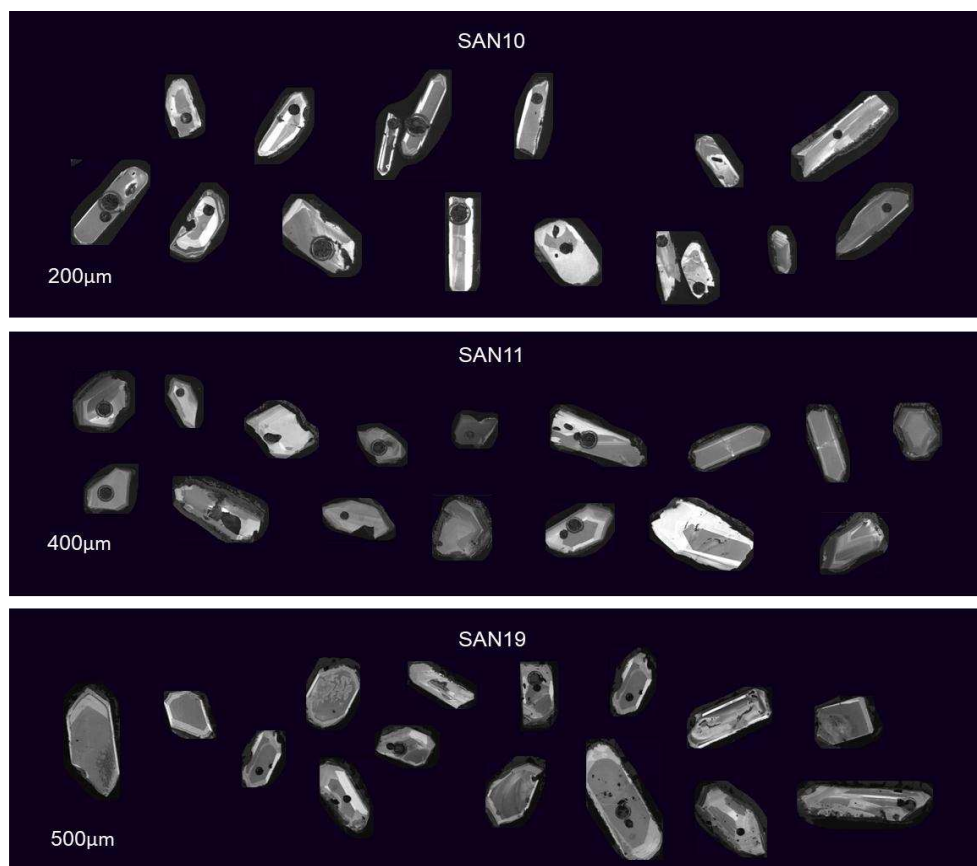


Figure 5.4 Cathodoluminescence micrographs of analyzed zircons from the selected Cenozoic intrusives of the Piura segment of the PCB. Laser spot size: 30µm.

The Paltashaco pluton records crystallization ages spanning the Paleocene to Oligocene (58.9 ± 0.6 to 29.1 ± 0.4 Ma). The Tucto–Penachi unit yielded Oligocene ages of 32.7 ± 0.3 and 29.6 ± 0.6 Ma. The Pamparumbe intrusion ranges from the Eocene to Miocene (43.6 ± 0.7 to 19.9 ± 0.4 Ma), while the Mallingas pluton is constrained to the Eocene with ages of 45.2 ± 0.3 and 40.7 ± 0.3 Ma. The Rumipite massif also records an Eocene age of 40.2 ± 0.4 Ma. Three smaller unnamed intrusions yielded ages of 35.9 ± 0.2 Ma (IG22-33; UNE), 20.2 ± 0.3 Ma (SAN08), and 20.2 ± 0.4 Ma (SAN09). The latter two intrusions (SAN08 and SAN09) are hereafter referred to as the UNW unit.

5.4.2 Whole rock geochemistry

The Cenozoic intrusions are dominated by felsic compositions ranging from granodiorite to granite (Figs. 5.5A, 5.6A), whereas the Mallingas unit comprises both felsic (granite) and mafic to intermediate lithologies (gabbro–diorite). All intrusions display patterns typical of convergent margin environments (Figs. 5.6G–H). Geochemical signatures are metaluminous to weakly peraluminous (Figs. 5.6C–D), consistent with magmas derived from crust–mantle interaction. Most samples plot within the magnesian field (Fig. 5.6E), indicating crystallization from relatively oxidized arc magmas, and are predominantly calcic to calc-alkalic (Fig. 5.6F), suggesting derivation from a subduction-related setting.

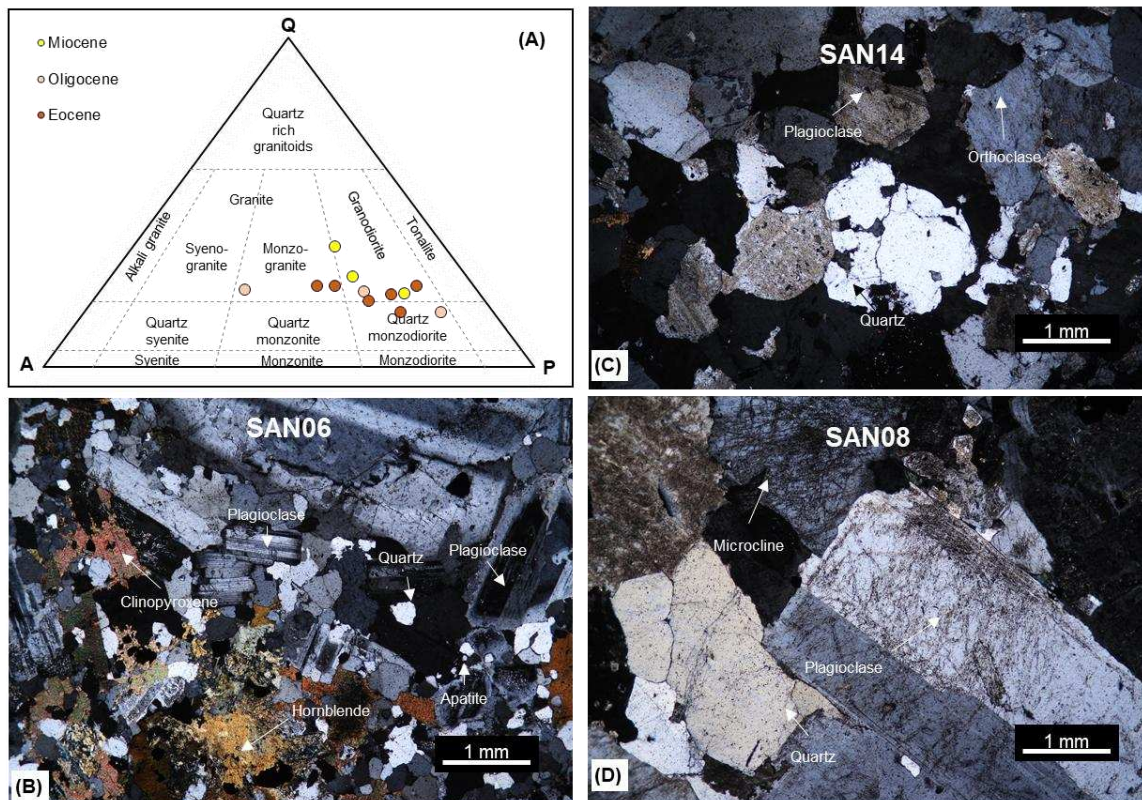


Figure 5.5 (A) QAP diagram used to classify intrusive igneous rocks based on their mineralogy. (B–D) Photomicrographs of Mesozoic rocks displaying main mineral assemblages.

Chondrite-normalized REE patterns (Fig. 5.6B) show moderate to pronounced enrichment in LREEs relative to HREEs, coupled with moderate to strong negative Eu anomalies. These anomalies reflect efficient plagioclase fractionation during magma evolution, except in the Tucto–Penachi unit, where the anomaly is absent.

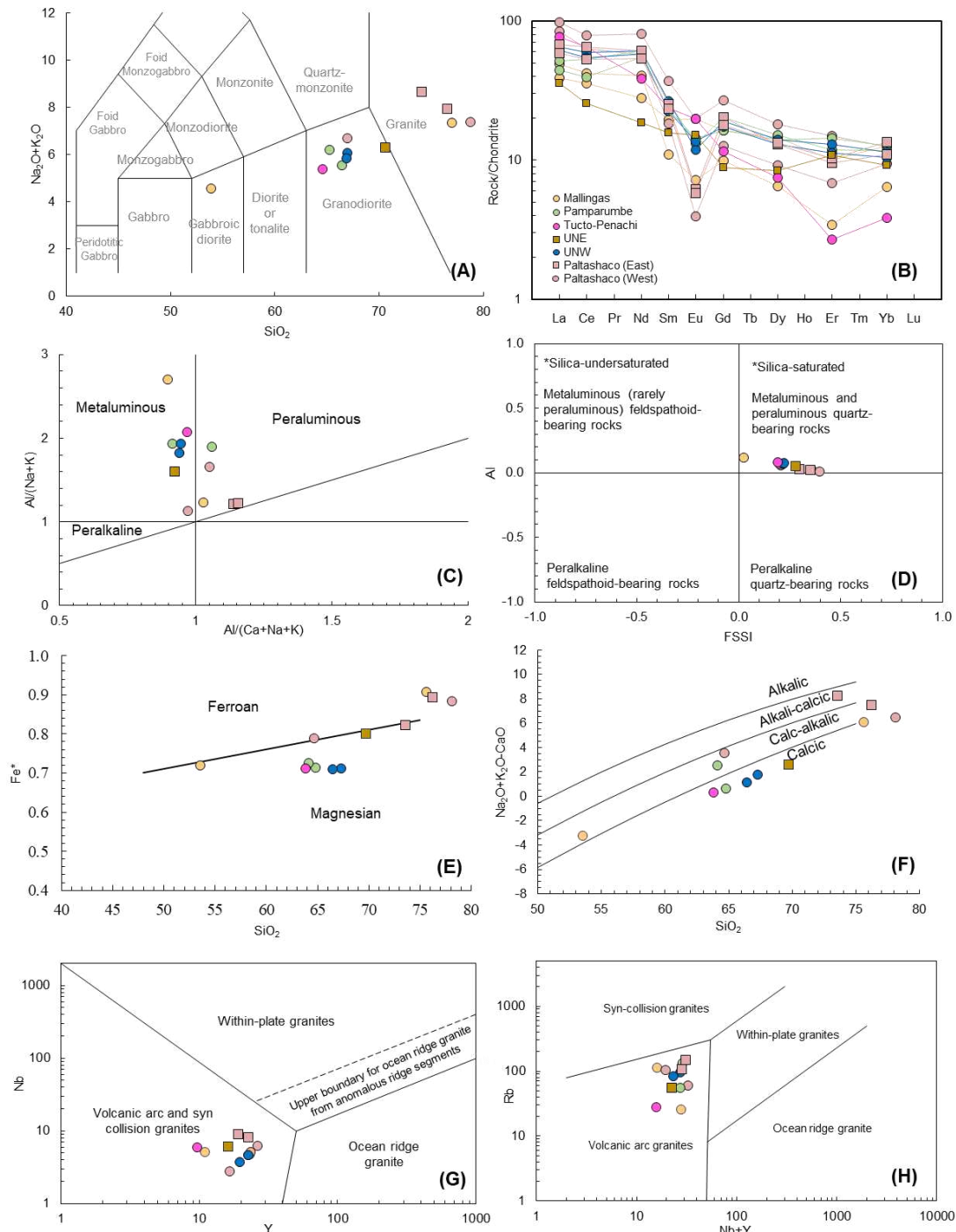


Figure 5.6 Whole rock classification diagrams. (A) Total alkali versus silica plot (Middlemost, 1994). (B) Chondrite-normalized rare element plot, normalizing values are from Sun and McDonough (1989). (C) Molar A/CNK versus molar A/NK plot. (D) Plot of alkalinity index (AI) vs feldspathoid silica-saturation index (FSSI). (E) SiO₂ versus Fe* plot (Frost et al. 2001). (F) Na₂O+K₂O-CaO as a function of SiO₂, fields are from Frost & Frost, 2008. (G-H) Tectonic discriminating plots from Pearce, J. (1982).

Petrographically, the rocks are phaneritic and exhibit hypidiomorphic granular textures with medium grain sizes (0.7–5 mm; Fig. 5). Mineral assemblages vary systematically among units: Mallingas (Qz 15%, Kfs 17%, Pl 58%, accessory Ap and Hbl); UNW (Qz 19–30%, Kfs 13–22%, Pl 37–53%, Bt 7–10%, accessory Ap); Pamparumbe (Qz 16–24%, Kfs 13–24%, Pl 40–44%, Ms <12%, Hbl <10%, Bt 8–12%, Cpx <5%); UNE (Qz 22%, Kfs 27%, Pl 42%, Ms <10%); Tucto–Penachi (Qz 18%, Kfs 18%, Pl 42%, Ms 12%, Bt 12%); and Paltashaco (Qz 16–21%, Kfs 10–42%, Pl 26–69%, Ms 11–18%) (Table S8). These modal proportions highlight the predominance of felsic quartz–feldspar–mica assemblages, with variable contributions of mafic phases. Secondary alteration is minor and limited to partial sericitization of micas, indicating the preservation of primary mineralogical and textural features.

5.4.3 Zircon isotopic $\epsilon\text{Hf}_{(t)}$ and $\delta^{18}\text{O}$ analyses

A total of 133 zircon isotopic analyses were obtained, including 97 $\epsilon\text{Hf}_{(t)}$ and 36 $\delta^{18}\text{O}$ measurements. The Pamparumbe and UNW units yield predominantly positive $\epsilon\text{Hf}_{(t)}$ values (+2.6 to +8.2; Fig. 5.7A) combined with mantle-like $\delta^{18}\text{O}$ signatures (4.8–6.4 ‰; Figs. 5.7B). These isotopic characteristics are consistent with derivation from depleted mantle sources with only minor assimilation of crustal material (Valley et al., 2005; Vervoort & Blichert-Toft, 1999). In contrast, the Rumipite pluton records distinctly negative $\epsilon\text{Hf}_{(t)}$ values (–5.2 to –1.4) coupled with elevated $\delta^{18}\text{O}$ values (5.7–8.1 ‰), suggesting derivation from evolved crustal sources with minimal mantle input.

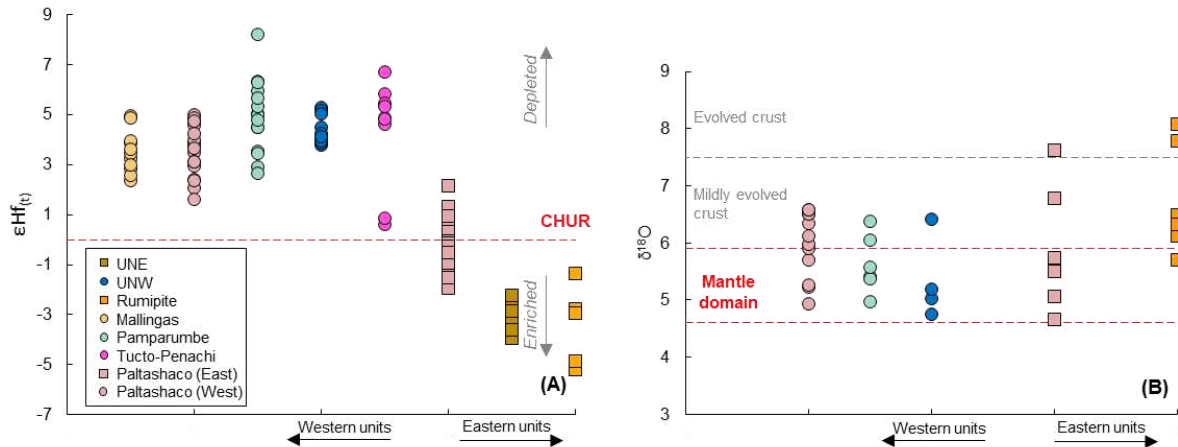


Figure 5.7 Isotopic data for intrusive samples from the Piura segment. (A) $\epsilon\text{Hf}_{(t)}$ values for individual intrusive bodies. (B) $\delta^{18}\text{O}$ values for individual intrusive bodies, with reference fields from Valley et al. (2005).

The Paltashaco unit displays variable isotopic signatures, with $\epsilon\text{Hf}_{(t)}$ ranging from –1.9 to +5.0 and $\delta^{18}\text{O}$ values from 4.7 to 7.7 ‰. Owing to its large extent, two isotopic domains can be distinguished separated by the Canchaque fault (Fig. 5.2). The western domain is characterized by more juvenile signatures ($\epsilon\text{Hf}_{(t)}$ +1.5 to +5.0; $\delta^{18}\text{O}$ 4.9–6.5 ‰), consistent with contributions from a depleted mantle

source. By contrast, the eastern domain yields lower ϵHf_t values (−1.9 to +2.2) and higher $\delta^{18}\text{O}$ values (4.7–7.6 ‰), reflecting enhanced crustal involvement in magma generation.

The Mallingas and Tucto–Penachi units also record positive ϵHf_t values (+0.8 to +6.7), indicating derivation from a depleted or hybrid source. In contrast, the UNE unit exhibits lower ϵHf_t values (−3.9 to −2.3), pointing to a more enriched source. However, the absence of $\delta^{18}\text{O}$ data for these intrusions currently precludes a more definitive interpretation of their genesis.

5.4.4 Zircon geochemistry

A total of 97 trace-element analyses were performed on transparent, inclusion-free zircon crystals ranging from 30 μm to 0.8 mm in size (Fig. 5.4). Zircons display trace-element patterns characteristic of arc magmatism, including overall enrichment in HREEs relative to LREEs and consistent negative Pr and Eu anomalies (Fig. 5.8A).

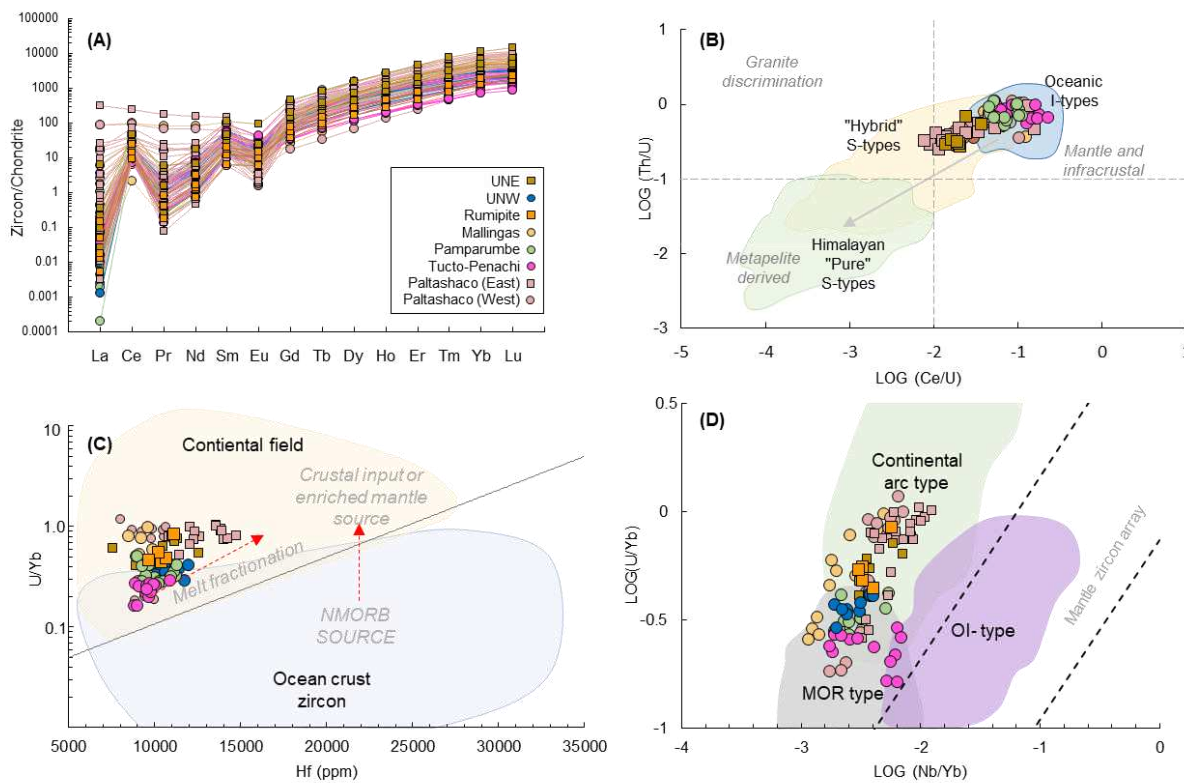


Figure 5.8 Zircon-based classification diagrams. (A) Chondrite-normalized rare element plot. (B) Th/U vs Ce/U plot defining I-type, S-type and hybrid S-type granitoids, boundaries are from Roberts et al. (2024). (C) U/Yb vs Hf, fields for continental and oceanic zircon are from Grimes et al., (2007) (D). Tectonic discrimination diagram, boundaries are from Grimes et al., (2015). MOR: Mid ocean ridge. OI: ocean island.

The Tucto–Penachi, Pamparumbe, UNW, and Mallingas units predominantly exhibit similar geochemical characteristics, with zircon compositions that show affinities to I-type granitoids and oceanic-like sources (Figs. 5.8B–D). A subset of zircons from Mallingas, however, displays more pronounced continental signatures, suggesting localized assimilation of crustal material. In contrast,

the Rumipite and UNE units are dominated by continental-type zircon compositions, with affinities to S-type granitoids and felsic crustal sources, and only minor overlap with the I-type field. The Paltashaco intrusion presents a mixed geochemical signature, with zircons recording both continental and oceanic affinities, consistent with hybrid S- and I-type granitoid sources.

Trace-element ratios provide further insights into magmatic sources and tectonic setting. Although the data display considerable scatter, some coherent trends can still be discerned. Zircon U/Yb and Th/U ratios are particularly diagnostic in distinguishing continental versus oceanic provenances (Grimes et al., 2007, 2015). Most samples yield U/Yb values between 0.3 and 1, typical of continental arcs (Fig. 5.9A). However, some zircons from the Mallingas, Paltashaco, and Pamparumbe intrusions extend to lower values approaching those of NMORB. An exception is the Tucto–Penachi unit, which consistently yields U/Yb values >0.3 , more typical of primitive sources. Th/U ratios generally fall within the felsic zircon field, although subsets of Paltashaco, Pamparumbe, and Tucto–Penachi record values >1 , indicating transitional magmas that incorporated both mafic and felsic components (Fig. 5.9B).

The La/Sm ratio, which reflects the degree of LREE enrichment and fractionation processes (Zhu et al., 2007), varies among the intrusions. Paltashaco and UNE display the highest values, indicating enhanced LREE enrichment that is consistent with either more evolved magmas or partial melts derived from a more enriched source. In contrast, Pamparumbe shows markedly lower ratios (Fig. 5.9C), suggesting more limited fractionation or derivation from a comparatively depleted source.

Lu/Hf, which is sensitive to differentiation trends and source fertility (Gardiner et al., 2018; Luffi & Ducea, 2022), follows a broadly similar pattern. Paltashaco and UNE yield the highest values, pointing to more pronounced differentiation and/or melting of a fertile lithology. The lower ratios in UNW and Tucto–Penachi (Fig. 5.9D) imply either less-evolved magmas or melting of a less fertile, possibly more refractory source.

The Yb/Gd ratio, commonly used to evaluate the role of residual garnet and depth of magma generation (Drabon et al., 2024; Lu et al., 2019). The low Yb/Gd values in UNW and Tucto–Penachi point to melting within the garnet stability field, suggesting a stronger lower-crust or lithospheric mantle contribution. In contrast, the remaining units display more dispersed values, with UNE and Paltashaco reaching the highest ratios (Fig. 5.9E). These elevated Yb/Gd values are consistent with reduced garnet influence, which aligns with their more fractionated and fertile geochemical signatures.

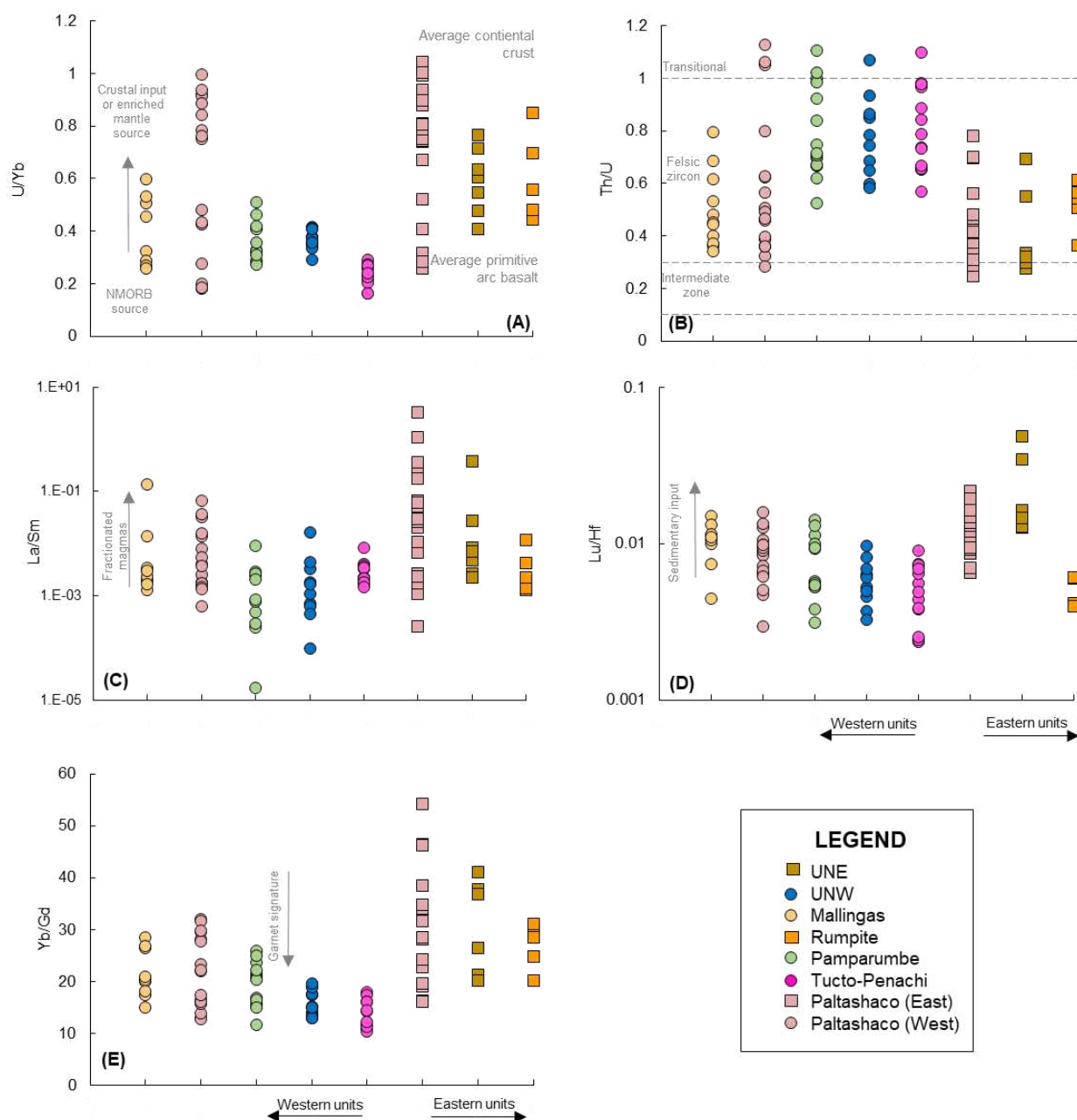


Figure 5.9 Zircon bivariate plots illustrating geochemical proxies per unit. (A–B) Melt maturity indicators with reference fields from Grimes et al. (2015): (A) U/Yb and (B) Th/U. (C) La/Sm. (D) Lu/Hf. (E) Yb/Gd.

5.4.5 Bouguer gravity anomalies

Bouguer gravity anomalies across the study area range from ca. +150 to –150 mGal (Fig. 5.1B). The highest positive values (+100 to +150 mGal) occur over the Illescas and Piura highs, consistent with dense crustal domains or mafic basement compositions (e.g., Dufr  chou et al., 2011). Moderate positive anomalies characterize the ATB and the Celica–Lancones domain (arc and its associated basin). In contrast, strongly negative anomalies (down to –150 mGal) in the northwestern Peruvian Cordillera indicate a lower density crustal basement. Localized moderate positive anomalies (up to ~30 mGal) within the Cordillera coincide spatially with exposures of the Rumipite diorite–tonalite

pluton, suggesting a local enhancement of the gravity field associated with these intermediate-mafic intrusive bodies, which possess relatively higher bulk density compared to the surrounding crust.

Within the low-elevation forearc region, the Sechura Basin exhibits a gradual west-to-east transition from positive to negative Bouguer values. Because the Bouguer correction removes the gravitational effects of surface topography and near-surface relief, these variations primarily reflect subsurface density contrasts rather than elevation differences (Szwilius et al., 2016; and references therein). The observed gradient, from positive anomalies in the west to increasingly negative anomalies eastward, implies a significant crustal boundary separating denser, potentially mafic basement in the west from less dense, more felsic continental crust to the east.

5.5 Discussion

5.5.1 Cenozoic geochronology of the Piura segment of the PCB

This study presents the most comprehensive U–Pb zircon geochronological dataset for the Piura segment of the PCB, establishing a more robust temporal framework than previous studies that relied primarily on K–Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ dating (Hama, 1990; Litherland, 1994; Ulrich, 2005; Winter, 2008) or on indirect stratigraphic and geographic correlations (De la Cruz Wetzell, 1995; Jaimes Salcedo, 2012; Reyes Rivera & Caldas Vidal, 1987; Wilson, 1984). Our results demonstrate that Cenozoic intrusions in this region were emplaced from Paleocene to Miocene, with magmatic activity beginning at least by 60 Ma.

The Paltashaco and Pamparumbe plutons, previously mapped broadly as Cretaceous to Paleogene units (Geological maps of the Las Lomas 10c-(0861), Ayabaca 10d-(0961), Chulucanas 11c-(0860), Morropón 11d-(0960), and Huancabamba 11e-(1060) quadrangles), are now tightly constrained by new U–Pb zircon geochronology. The Paltashaco pluton yields Paleocene to Oligocene ages (59–29 Ma), whereas the Pamparumbe pluton records younger, Eocene to early Miocene ages (44–20 Ma).

The dataset also provides the first radiometric ages for the Mallingas intrusive body (45–41 Ma; Eocene) and for two previously undifferentiated intrusions, UNE (ca. 36 Ma) and UNW (ca. 20 Ma), which had been tentatively assigned to the Paleogene based solely on field relationships (Jaimes Salcedo et al., 2013; Reyes Rivera & Caldas Vidal, 1987).

Our new data also significantly narrow the age of the Rumipite pluton, now assigned to ca. 40 Ma, compared to its previously broad range of 83–16 Ma based on K–Ar method (Hama, 1990). Finally, we confirm an Oligocene age of 33–30 Ma for the Tucto–Penachi unit, as previously constrained by CGS_INGEMMET (2017).

5.5.2 Origin of the magmas

Cenozoic intrusions exhibit metaluminous to peraluminous compositions, predominantly magnesian character, and calcic to alkali-calcic affinities (Figs. 5.6C–F), consistent with crystallization in an active continental margin setting (Frost & Frost, 2008; Hoskin & Schaltegger, 2003; Pearce, 1982). Zircon trace-element patterns show an enrichment of HREEs relative to LREEs, which supports continental arc magmatism (Fig. 5.8A).

The Eocene Mallingas intrusion records bimodal magmatism ranging from granite to gabbro–diorite compositions (Fig. 5.6A), reflecting the coexistence of felsic and mafic magmas within an active continental margin. Zircon geochemistry reveals a predominant oceanic I-type affinity, with a minor population showing more evolved, continental signatures (Fig. 5.8). The moderate degree of magmatic differentiation, coupled with trace-element ratios (La/Sm, Lu/Hf; Drabon et al., 2024; Gardiner et al., 2018; Luffi & Ducea, 2022) and positive $\epsilon\text{Hf}_{(t)}$ values (+2.4 to +5.0), suggests that the Mallingas magmas were sourced from a depleted mantle that experienced limited crustal assimilation. Such signatures are consistent with a mantle wedge modified by slab-derived fluids. Comparable isotopic and geochemical patterns have been reported farther north in southwestern Ecuador (Witt & Bourgois, 2010) (Iglesias et al., in review).

The Eocene–early Miocene Pamparumbe pluton consists mainly of granodiorite, tonalite, and diorite (Reyes Rivera & Caldas Vidal, 1987; this study). Its hornblende- and clinopyroxene-bearing assemblages indicate hydrous arc magmatism operating at mid- to lower-crustal levels (Chang et al., 2021; Liu et al., 2021; Putirka, 2008). Zircon compositions show oceanic I-type affinities (Fig. 5.8), low U/Yb (Fig. 9A), and Th/U ratios spanning felsic-like to MORB-like values (Fig. 5.9B). Furthermore, high $\epsilon\text{Hf}_{(t)}$ values (+2.7 to +6.6; Fig. 5.7), and mantle-like $\delta^{18}\text{O}$ values ($5.3\text{‰} \pm 0.6\text{‰}$; Roberts & Spencer, 2015) support derivation from a depleted mantle source with minimal crustal contamination.

The Oligocene Tucto–Penachi unit is composed predominantly of granodiorite, with subordinate tonalite, quartz diorite, and granite (Jaimes Salcedo et al., 2013; this study). Geochemically, this unit differs from the other intrusions by lacking a pronounced negative Eu anomaly (Figs. 5.6B and 5.8A) and by displaying the lowest U/Yb, Lu/Hf, and Yb/Gd ratios among the studied plutons (Fig. 5.9). These features indicate crystallization from relatively primitive magmas, under conditions consistent with garnet stability (Bell & Kirkpatrick, 2021; Drabon et al., 2024). Zircons plot mainly within the oceanic I-type field (Fig. 5.8) and yield positive $\epsilon\text{Hf}_{(t)}$ values ranging from +4.6 to +6.7, with a few slightly lower values near +1, suggesting derivation from a depleted mantle source with minor crustal input. Overall, the geochemical and isotopic characteristics of the Tucto–Penachi unit point to a

hybrid but dominantly mantle-derived origin. Nevertheless, the lack of $\delta^{18}\text{O}$ data currently limits a more precise description.

The Eocene Rumipite unit consists of tonalites and diorites containing abundant hornblende phenocrysts (Reyes Rivera & Caldas Vidal, 1987). This unit represents the most strongly crustal-derived intrusion within the suite. Zircons plot exclusively within the S-type granitoid field (Fig. 5.8) and exhibit U/Yb (Fig. 9A) and Th/U (Fig. 5.9B) ratios, consistent with derivation from evolved felsic sources. The consistently negative $\epsilon\text{Hf}_{\text{(t)}}$ values (-4.8 to -1.4 ; Fig. 5.7A), together with elevated $\delta^{18}\text{O}$ values (5.7 – 8‰ ; Fig. 5.7B), indicate melting of continental crust with negligible mantle contribution.

The Miocene UNW granitoid ranges from granite to granodiorite (Fig. 5.5A) and displays zircon compositions with oceanic-like signatures (Fig. 5.8). Low U/Yb ratios approaching N-MORB values (Fig. 5.9A), together with relatively high Th/U (Fig. 9B), suggest limited differentiation and minimal crustal interaction. The consistently intermediate - high $\epsilon\text{Hf}_{\text{(t)}}$ values ($+3.9$ to $+5.3$) and mantle-like $\delta^{18}\text{O}$ compositions (4.7 – 6.4‰ ; Fig. 5.7) indicate that the UNW magmas were dominantly, generated from a depleted mantle source.

The Paleocene–Oligocene Paltashaco pluton, the largest intrusive body in the region, exhibits pronounced geochemical and isotopic zonation between its western and eastern domains, separated by the Canchaque fault (Fig. 5.2). The western domain is characterized by oceanic I-type zircon signatures (Fig. 5.8), low U/Yb and Th/U ratios, and intermediate to low La/Sm and Lu/Hf, values, consistent with limited crustal involvement. Positive $\epsilon\text{Hf}_{\text{(t)}}$ values (up to $+5$) and mantle-like $\delta^{18}\text{O}$ compositions (4.9 – 6.5‰ ; Fig. 5.7) further support derivation from a depleted mantle source. In contrast, the eastern domain is dominated by zircon populations with S-type affinities, higher La/Sm and Yb/Gd ratios, lower $\epsilon\text{Hf}_{\text{(t)}}$ (-1.9 to $+2.2$), and $\delta^{18}\text{O}$ values reaching 7.6‰ , reflecting a significant contribution of crustal components to the magma source.

The spatial distribution of geochemical and isotopic data reveals a clear west–east compositional contrast, which is broadly delineated by the Canchaque Fault (Fig. 5.10). Intrusions located west of the fault (Mallingas, Pamparumbe, Tucto–Penachi, UNW, and the western Paltashaco domain) display I-type affinities, low U/Yb and Lu/Hf ratios, high Th/U, and mantle-like isotopic compositions, indicating derivation from depleted mantle sources with limited crustal assimilation. In contrast, intrusions to the east (Rumipite, UNE, and the eastern Paltashaco domain) are dominated by S-type affinities, elevated La/Sm and Lu/Hf ratios, and evolved isotopic signatures, including low $\epsilon\text{Hf}_{\text{(t)}}$ and high $\delta^{18}\text{O}$ values (Fig. 5.7), consistent with partial melting of continental crust. These patterns suggest that the Canchaque fault represents a lithospheric-scale boundary separating an oceanic-affinity basement in the west (Fig. 5.10) from a continental block in the east. This

interpretation is consistent with structural boundaries proposed further north in southern Ecuador (Iglesias et al., in review).

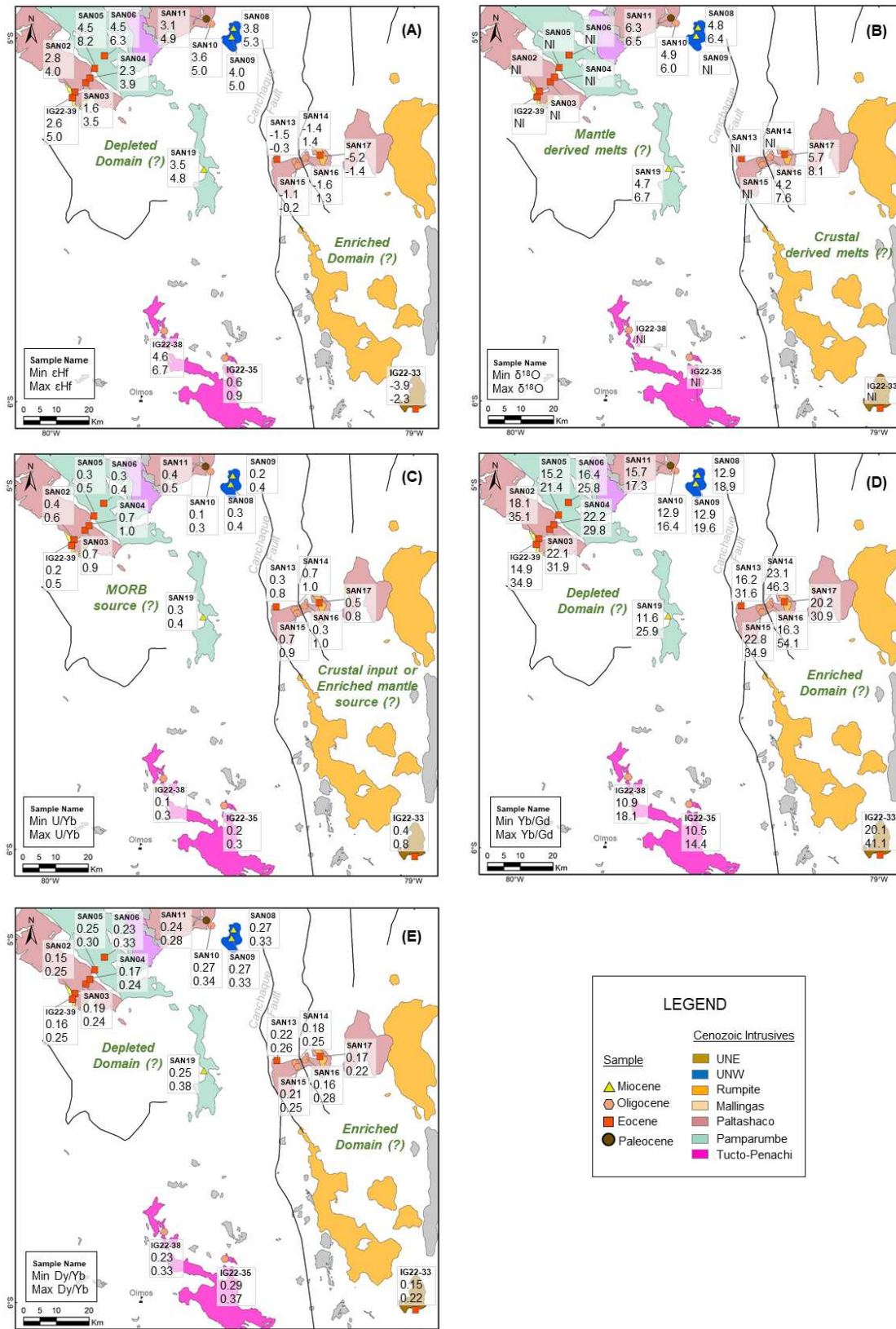


Figure 5.10 Spatial distribution of isotopic and trace-element compositions of studied samples, illustrating heterogeneity associated with the Canchaque Fault. (A) ϵ_{Hf} . (B) $\delta^{18}\text{O}$. (C) U/Yb ratio. (D) Yb/Gd ratio. (E) Dy/Yb.

The scatter observed in trace-element ratios (Fig. 5.9) likely reflects not only spatial heterogeneity but also temporal variations in tectonic regime. Periodic shifts between compression and extension, documented in the structural and stratigraphic record of the region (Espurt et al., 2018; Hungerbühler et al., 2002; Jaillard et al., 2000, 2025), would have modulated crustal thickness and magma ascent pathways, influencing the degree of assimilation, fractionation, and residual mineral stability. Therefore, studies that emphasize temporal control over geochemical variations (i.e., George et al., 2021; Witt et al., 2023) are not contradictory but complementary, as temporal processes likely acted as secondary modifiers superimposed on the primary spatial control exerted by lithospheric architecture.

5.5.3 Basement architecture of northern Peru

Positive Bouguer anomalies along the western segment of the study area (up to +150 mGal; Fig. 5.1B) indicate the presence of denser crust characteristic of an oceanic-affinity basement composed of mafic units (Szwilius et al., 2016). In contrast, markedly negative anomalies to the east reflect lower-density continental crust (Hawkesworth & Kemp, 2006a). Because Bouguer anomalies incorporate topographic corrections, removing the effects of the higher elevations of the Peruvian Western Cordillera to the east and the lower elevations of the forearc region to the west, the observed west–east contrast cannot be attributed to surface relief (Fig. 5.1B). Instead, the anomaly gradient reflects fundamental differences in basement architecture.

Geochemical and isotopic data independently support this architectural boundary (see section 5.5.2). Intrusions west of the Canchaque Fault exhibit juvenile, mantle-like signatures, whereas eastern intrusions display evolved zircon ϵHf_t and elevated $\delta^{18}\text{O}$ values consistent with continental crustal input. The spatial correspondence between gravity data and isotopic compositions therefore indicates the presence of two distinct basement domains, oceanic to the west and continental to the east, which exerted a primary control on magma chemistry during Cenozoic magmatism.

5.5.4 Spatio-Temporal model for Cenozoic magmatism in north Peru

The onset of Cenozoic magmatism is recorded by the western Paltashaco pluton (~60 Ma), emplaced during a period of high convergence rates and more orthogonal subduction (N to NE; Espurt et al., 2018; Jaillard, 2025; Maloney et al., 2013; Seton et al., 2012). The compressional setting induced crustal thickening (Jaillard & Soler, 1996; Witt et al., 2023), as reflected in zircon trace-element signatures characterized by low Yb/Gd and high Dy/Yb ratios (Figs. 5.10D–E), consistent with melting in the presence of residual garnet at deeper crustal levels. The western Paltashaco unit also exhibits juvenile isotopic compositions (ϵHf_t up to +5.0 and mantle-like $\delta^{18}\text{O}$ values; Fig. 5.7),

indicative of derivation from an oceanic-affinity basement. We therefore infer that these magmas were generated at depths sufficient to restrict upper-crustal interaction, allowing them to retain the geochemical and isotopic characteristics of their basement source (Fig. 5.11A).

A major shift to an extensional regime occurred during the middle Eocene (~45–40 Ma), likely driven by slab retreat (Espurt et al., 2018; Jaillard et al., 2000; Witt et al., 2023). This transition coincided with a peak in magmatic activity, marked by emplacement of the Mallingas (45–41 Ma), Rumipite (~40 Ma), and Pamparumbe (~44 Ma) plutons. In the western domain (Mallingas and Pamparumbe), extension likely facilitated the ascent of mantle-derived melts. Zircon TE data from this region show increasing Yb/Gd ratios (Fig. 5.10D) and a general trend toward lower Dy/Yb ratios (partially overlapping with the ~60 Ma population; Fig. 5.10E), suggesting relative crustal thinning. In contrast, the eastern Rumipite pluton exhibits elevated $\delta^{18}\text{O}$ values and negative $\varepsilon\text{Hf}_{\text{(t)}}$ signatures, indicating substantial assimilation and/or partial melting of evolved crustal material, a pattern well documented in other extensional settings (i.e., Foley et al., 2023; Ódri et al., 2020). These contrasting signatures demonstrate that extension did not generate a uniform magmatic response but instead produced melts whose chemistry was strongly modulated by the underlying basement composition (Fig. 5.11B).

A compressional regime was reestablished from the late Eocene to middle Oligocene (~40–25 Ma), driven by an increase in subduction rates and a shift to nearly E–W convergence following the breakup of the Farallon Plate (Jaillard, 2025; Maloney et al., 2013; Pardo-Casas & Molnar, 1987; Seton et al., 2012). This interval was characterized by crustal thickening, thrust-driven exhumation (Espurt et al., 2018), and widespread magmatism, represented by the UNE (~36 Ma), eastern Paltashaco (34–33 Ma), and Tucto–Penachi (33–30 Ma) plutons. In the eastern domain, zircon TE compositions from the UNE and eastern Paltashaco plutons plot entirely within the continental crust field (Figs. 8C–D), supported by $\delta^{18}\text{O}$ values of 6.7–7.8‰ and $\varepsilon\text{Hf}_{\text{(t)}}$ as low as –4 (Fig. 5.7). However, mantle-like $\delta^{18}\text{O}$ values and more radiogenic $\varepsilon\text{Hf}_{\text{(t)}}$ up to +2.2 in the eastern Paltashaco pluton reveal a minor but significant mantle-derived contribution, indicating a hybrid source. A comparable hybrid signature is present in the western domain: zircon $\varepsilon\text{Hf}_{\text{(t)}}$ values from the Tucto–Penachi pluton define two clusters—a depleted, juvenile mantle-like group (+4.6 to +6.7) and a less radiogenic group (~+1) reflecting partial assimilation of continental crust. We propose that crustal thickening during this compressional phase enhanced magma mixing and assimilation, locally blurring the otherwise well-defined compositional contrast between the eastern continental and western oceanic-affinity basement domains (Fig. 5.11C).

The early Miocene was marked by a reduction in convergence rates (Maloney et al., 2013), likely inducing a renewed extensional regime between 22 and 19 Ma. This tectonic shift resulted in

diminished thrusting, with internal shortening of less than ~6 km (Espurt et al., 2018), and promoted magmatism in the western domain. The emplacement of the Pamparumbe and UNW plutons at ~20 Ma represents this magmatic phase. Isotopic data from these units, including high $\epsilon\text{Hf}(t)$ values coupled with mantle-like $\delta^{18}\text{O}$ signatures, indicate clear inheritance from an oceanic-type basement (Figs. 5.8C–D), suggesting limited crustal assimilation during magma ascent.

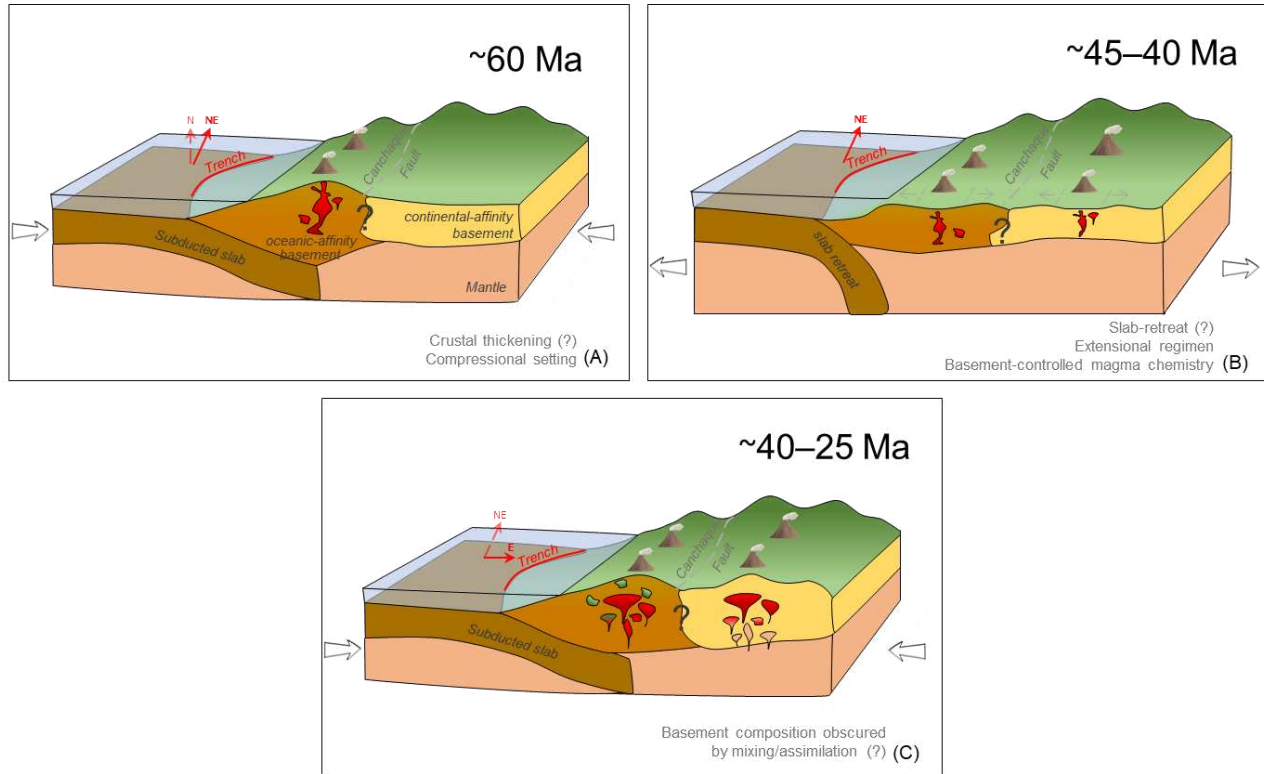


Figure 5.11 Simplified cartoon illustrating the tectono-magmatic evolution of northern Peru from 60 to 25 Ma. (A) ~60 Ma; (B) 45–40 Ma; (C) 40–25 Ma.

5.6 Conclusions

This study presents the first comprehensive U–Pb zircon geochronological framework for Cenozoic magmatism in northern Peru, documenting a prolonged magmatic history from ~60 to 20 Ma. The new age constraints refine the emplacement chronology of major intrusive units, including the Paltashaco (59–29 Ma), Pamparumbe (44–20 Ma), and Rumipite (~40 Ma) plutons, as well as the undifferentiated UNE (36 Ma) and UNW (20 Ma) units.

The compositional evolution of Cenozoic magmatism in northern Peru is governed primarily by the heterogeneous architecture of the underlying lithosphere. To the west, the basement exhibits oceanic affinity, characterized by positive $\epsilon\text{Hf}(t)$, mantle-like $\delta^{18}\text{O}$ values, and positive Bouguer anomalies. In contrast, the eastern basement displays continental affinity, defined by negative $\epsilon\text{Hf}(t)$, elevated $\delta^{18}\text{O}$ signatures, and negative Bouguer anomalies. Although the transition between these domains is

not sharply observed in Bouguer anomaly map, the integrated geochemical and isotopic data allow us to tentatively place the boundary along the Canchaque Fault.

Plutonic bodies west of the Fault (Mallinas, Pamparumbe, Tucto–Penachi, UNW, and the western Paltashaco intrusions) consistently record depleted, mantle-like isotopic signatures. Eastern plutons (Rumipite, UNE, and the eastern Paltashaco intrusions) instead display enriched isotopic compositions, indicating derivation from, or significant interaction with, continental crust.

Tectonic regime changes exert a secondary but important influence by modulating mantle–crust interactions, melt generation depths, and magma ascent pathways. Periods of compression promoted crustal thickening and enhanced magma mixing, slightly obscuring the lithospheric compositional contrast. In contrast, extensional phases facilitated more efficient melt ascent and helped preserve the distinct geochemical signatures imposed by the underlying basement.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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CHAPTER 6: UNRAVELLING THE COOLING AND EXHUMATION HISTORY OF SOUTHERN ECUADOR: APATITE FISSION-TRACK INSIGHTS FROM A COMPLEX ANDEAN SEGMENT

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Declaration statement

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ABSTRACT

The cooling and exhumation history of orogens along subduction systems provides unique constraints on regional tectonic evolution, reflecting the interplay between crustal deformation, fault activity, and plate convergence dynamics. Southern Ecuador lies within a transitional zone between the northern Andes, where collisional and accretionary events have played a dominant tectonic role, and the central Andes, where such processes are largely absent. This position offers a unique context to evaluate whether exhumation in southern Ecuador reflects the influence of northern Andean accretionary events, changes in subduction dynamics, or local tectonic processes. We present new apatite fission-

track thermochronology data from seventeen plutonic rock samples across five crustal blocks in southern Ecuador: Western Cordillera, Eastern Cordillera, Intermontane basins, Celica-Lancones, and Amotape-Tahuin. Our results show that cooling ages are internally consistent within each block but differ across them, revealing a pattern of diachronous exhumation in most blocks. Some blocks instead record pronounced post-magmatic cooling, suggesting shallow pluton emplacement. Thermal history modeling, based on good-fit time–temperature paths indicates that regional exhumation initiated in the middle Eocene and persisted thereafter. The cooling trajectories group into three distinct, albeit partially overlapping, intervals: ~45–38 Ma (middle - late Eocene), ~40–30 Ma (late Eocene – early Oligocene), and ~33–25 Ma (Oligocene). We interpret this protracted and spatially variable exhumation as reflecting changes in the boundary conditions of the subduction system, including a transition from oblique to more orthogonal convergence and increases in convergence rates during the Cenozoic. However, the observed spatial variability is more directly linked to the progressive activation of major fault systems and, locally, to magmatic activity.

Key-words: Exhumation, cooling, Thermal-modelling, Tectonics, Convergence.

6.1 Introduction

The timing, drivers, and spatial expression of exhumation and topographic growth in the Andean mountain range remain intensively debated. Geological and thermochronology evidence demonstrates a long history of crustal deformation and exhumation since at least the Late Cretaceous (e.g., Bayona et al., 2008; Bermúdez et al., 2019; Boschman, 2021; Mora et al., 2006; Noriega-Londoño et al., 2020; Restrepo-Moreno et al., 2019; Ronda et al., 2022; Sandoval-Espinel et al., 2024; Spikings et al., 2005; Zapata et al., 2019). At the same time, multiple lines of evidence, including provenance analysis, biostratigraphic constraints and paleoenvironmental records, suggest that the most significant phase of Andean topographic growth occurred during the Miocene (e.g., Anderson et al., 2016; Echeverri et al., 2015; Giambiagi et al., 2022; Howlett et al., 2025; Mora et al., 2008; Mosolf et al., 2011; Steinmann et al., 1999), when accelerated uplift and crustal shortening established the Andes as a major orographic barrier (Horton, 2018).

The timing and mechanisms of exhumation and uplift are not uniform along the orogen and remain poorly constrained in many regions, particularly in Ecuador. Here, exhumation has been primarily attributed to successive oceanic terrane-accretion events, including the incorporation of Caribbean-derived terranes (e.g., Ruiz et al., 2004; Spikings et al., 2001; 2005; 2010), as well as to the subduction of aseismic ridges such as the Carnegie Ridge (Brichau et al., 2021; George et al., 2022; Gutscher, 2002; Margirier et al., 2023; Michaud et al., 2009; Spikings et al., 2010). To a lesser extent, exhumation has also been linked to changes in subduction dynamics, including variations in

convergence rates (Margirier et al., 2023; Spikings et al., 2005, 2010) and shifts in the subduction vector toward more orthogonal convergence (Jaillard, 2025).

Most thermochronological studies in Ecuador have focused on the northern segment of the margin (e.g., Bricchau et al., 2021; Margirier et al., 2023; Ruiz et al., 2004; Spikings et al., 2000, 2005, 2010), particularly between $\sim 1^{\circ}\text{N}$ and 2°S , where the Carnegie Ridge is currently subducting (Gutscher et al., 2000; Michaud et al., 2009). In contrast, southern Ecuador remains comparatively under-studied, with existing data largely limited to the Western and Eastern Cordilleras and the Amotape–Tahuin Block (ATB; Margirier et al., 2023; Spikings et al., 2001, 2005, 2010). Other structural blocks, including the Celica–Lancones Basin, the Intermontane basins, and the Inter-Andean Valley, have received little to no thermochronological attention (e.g., Espurt et al., 2018; Steinmann et al., 1999).

Southern Ecuador offers a strategic location to determine whether exhumation arises from terrane-accretion processes, shifts in subduction dynamics, or a combination of both. This region lies south of the Colombian–north central Ecuadorian margin, where accretionary events were pronounced and are inferred to diminish southward (Restrepo-Moreno et al., 2019; Ruiz et al., 2004; Spikings et al., 2000).

In this study, we present seventeen new AFT ages from bedrock samples from across southern Ecuador (Fig. 1B), spanning major crustal blocks including the Celica-Lancones Arc, the Western and Eastern Cordilleras, the ATB, and several Cenozoic intrusives intruding the Intermontane basins. Using thermal history modeling, we assess the timing and rates of exhumation and investigate spatial variations across strike. Our results indicate a protracted exhumation history punctuated by three cooling intervals. These intervals broadly correlate with changes in subduction dynamics, including transitions from oblique to more orthogonal convergence and increases in convergence rates. However, the distinct timings and rates of cooling between adjacent crustal blocks suggest that deformation was primarily accommodated along major crustal faults. Additionally, the temporal overlap between exhumation and late Oligocene–early Miocene arc magmatism suggests that magmatic processes may have contributed to enhanced exhumation in some segments of the study area.

6.2 Geological background and previous thermochronological studies

6.2.1 Geological overview

The Ecuadorian Andes represent a composite orogenic belt characterized as a bivergent orogen, with active thrusting along the western flank of the Western Cordillera and the sub-Andean ranges to the east (Fig. 6.1; Baby et al., 2013; Eguez et al., 2003; Hernández et al., 2024; Jaillard et al., 2005).

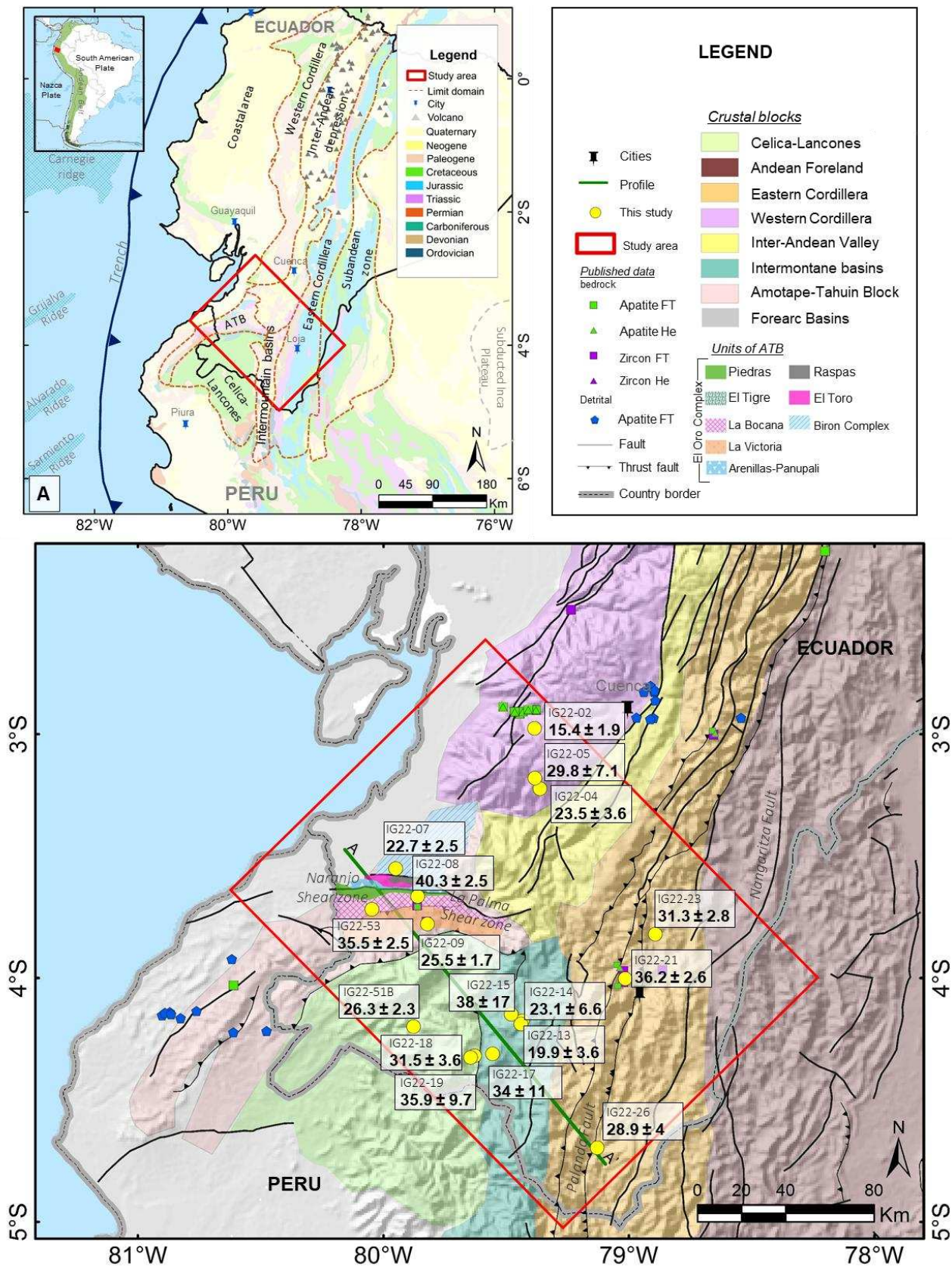


Figure 6.1 (A) Regional geology of central-southern Ecuador, highlighting major faults and crustal domains. (B) Sketch map of southern Ecuador showing intrusive rocks and sample locations analyzed in this study, along with published thermochronological data (ZFT, AFT, AHe, and ZHe) from Spikings et al. (2001, 2005, 2010); Espurt et al. (2018); Steinmann et al. (1999); and Margirier et al. (2023).

This orogeny has evolved through a complex history of subduction processes, involving Mesozoic-Cenozoic accretionary events (Aizprua et al., 2019; Hughes & Pilatasig, 2002; Jaillard, 2022; Kerr & Tarney, 2005; Reynaud et al., 1999; Vallejo et al., 2006; Witt et al., 2019), transitions between compressional and extensional settings (George et al., 2021; Witt et al., 2023) and variations in convergence rates (Maloney et al., 2013).

This intricate tectonic evolution has given rise to a number of distinct crustal blocks throughout southern Ecuador (Fig. 1A). Each region reflects a different geological history and lithological composition. To the east lies the Eastern Cordillera, also known as the Cordillera Real, which comprises Paleozoic to Lower Cretaceous metamorphic rocks (Kennan, 2000; Litherland, 1994). To the west of this range lies the Inter-Andean Valley, a structural depression filled with Cenozoic volcanic and sedimentary rocks (e.g., Hungerbühler et al., 2002). Adjacent to the west are the Intermontane basins, which were accompanied by sustained magmatic activity until ~5 Ma (e.g., Schütte et al., 2010). Further west lies the Cretaceous Celica-Lancones Basin, whose eastern segment is dominated by continental arc-related Late Cretaceous igneous rocks and volcanoclastic deposits (e.g., The Alamor and Celica Fms; Jaillard et al., 1996; Reynaud et al., 1996; Valarezo et al., 2019), while the central-western part is mainly composed of Late Cretaceous - Paleocene siliciclastic deposits (Jaillard et al., 1999; Winter et al., 2010; and references therein). Further west lies the ATB (Fig. 6.1B), an E-W trending low elevation range that shifts to a NE-SW orientation in Peru (Mourier, 1988; Witt et al., 2017). It can be subdivided into two main domains: the El Oro Complex, located south of the Naranjo Shear Zone (NSZ), which includes the Las Piedras Unit and the Tahuin Group (Noble et al., 1997; Riel et al., 2014), and a regional-scale *mélange* zone north of the NSZ, containing high-pressure rocks primarily from the Rapas Ophiolitic Complex and the Biron Complex (Riel et al., 2014).

6.2.2 Previous thermochronological studies

Published thermochronological data from southern Ecuador (Fig. 6.1B) are concentrated in the Western and Eastern Cordilleras and the ATB. These studies document three principal cooling phases at ~75–55 Ma (ZFT and AFT; Spikings et al., 2001, 2005), ~43–30 Ma (ZFT, AFT, and AHe; Spikings et al., 2010), and ~20–5 Ma (AFT, AHe; Margirier et al., 2023; Spikings et al., 2001, 2010), each linked to major tectonic and geodynamic events.

The ~75–55 Ma cooling phase in south Ecuador has been tentatively linked to the collision and accretion of the Pallatanga Terrane in north-central Ecuador (e.g., Spikings et al., 2001; 2005; 2010) or to increment in convergence rates (see Espurt et al., 2018). The ~43–30 Ma interval was initially

attributed to the collision of the Macuchi intra-oceanic arc with the Pallatanga Terrane (e.g., Spikings et al., 2001; 2005), but, more recently, this event has been linked exclusively to changes in plate kinematics (accelerated subduction rates; Espurt et al., 2018; Spikings et al., 2010). Cooling during the ~20–5 Ma interval has been widely associated with the onset of Carnegie Ridge subduction. For example, Spikings et al. (2000; 2001) proposed an early onset of Carnegie Ridge subduction at ~15–10 Ma based on the exhumation history of the Eastern Cordillera. In contrast, studies conducted in the Western Cordillera suggest a later onset of about 6–5 Ma (e.g., Margirier et al., 2023).

6.3 Sampling and methods

We collected seventeen bedrock samples from the Eastern Cordillera, the Western Cordillera, the Intermountain basins, the Celica-Lancones arc, and the ATB. Their locations and key details are provided in Table 6.1 and illustrated in Figure 6.1B.

Table 6.1 Sample locations and their zircon U-Pb ages

Sample Name	Lat.	Long.	Elevation (m)	U-Pb age (Ma)	$\pm 1\sigma$ (Ma)	Crustal block	Rock type	Source
IG22-02	-2.9769	-79.3842	3480.3	13.3	0.2	Western Cordillera	Granodiorite	Iglesias et al. (2025, submitted)
IG22-04	-3.2238	-79.3621	2246.37	29.7	2.0	Western Cordillera	Quartz-diorite	Iglesias et al. (2025, submitted)
IG22-05	-3.1809	-79.3824	2716.3	1035.7*	5.1	Western Cordillera	Granodiorite	Iglesias et al. (2025, submitted)
IG22-07	-3.5515	-79.9495	72.58	NA	NA	ATB (Biron complex)	Plutonic	This study
IG22-08	-3.6657	-79.8601	226.14	NA	NA	ATB (El Oro Complex)	Granodiorite	This study
IG22-09	-3.7678	-79.8249	749.56	200–1000		ATB (El Oro Complex)	Granite	This study
IG22-13	-4.1758	-79.4336	2024.37	18.5	0.2	Intermountain basins	Quartz-diorite	Iglesias et al. (2025, submitted)
IG22-14	-4.1891	-79.4395	2133.67	21.7	0.4	Intermountain basins	Tonalite	Iglesias et al. (2025, submitted)
IG22-15	-4.1486	-79.4793	1427.23	36.7	0.4	Intermountain basins	Tonalite	Iglesias et al. (2025, submitted)
IG22-17	-4.3109	-79.5552	1906.31	NA	NA	Intermountain basins	Plutonic	This study
IG22-18	-4.3300	-79.6262	1880.65	96.3	2.1	Celica-Lancones arc	Gabbro	This study
IG22-19	-4.3260	-79.5844	1965.76	94.7	0.9	Celica-Lancones arc	Granodiorite	This study
IG22-21	-4.0027	-79.0148	1546.22	NA	NA	Eastern Cordillera (Sabanilla)	Granodiorite	This study
IG22-23	-3.8204	-78.8921	874.21	191.6	1.8	Eastern Cordillera (Zamora)	Granodiorite	This study
IG22-26	-4.6966	-79.1287	1042.43	179.2	1.1	Eastern Cordillera (Zamora)	Granodiorite	This study
IG22-51B	-4.1992	-79.8767	549.1	100.8	1.8	Celica-Lancones arc	Diorite	This study
IG22-53	-3.7183	-80.0464	521.3	235.5	3.7	ATB (El Oro Complex)	Granite	This study

NA: not available. *: Inherited component, no crystallization age is available.

6.3.1 Apatite fission-track thermochronology

Apatite fission track thermochronology is based on the defects produced by the spontaneous fission of ^{238}U (Gleadow, 1981), which can be retained, or annealed depending on the temperature. Thus, this method can be used to reconstruct the thermal history of rocks by providing insights into their

cooling within the shallow crust (e.g., Armstrong et al., 2003; Gallagher et al., 1998; Gleadow et al., 2002). The temperature range in which fission tracks partially anneal is referred to as the partial annealing zone (PAZ), and is typically used to record the cooling of rocks within the 60°C to 120°C interval (Gallagher et al., 1998).

Sample preparation took place at the Pedagogical and Technological University of Colombia (UPTC), following Kohn et al. (2019) standard techniques. Samples were crushed, sieved, and separated using hydro-gravimetric and magnetic methods (Frantz magnetic separator with current increments from 0.2 to 1.2 Amperes) to isolate the heavy and non-magnetic fractions. Subsequently, gravimetric separation was carried out using standard heavy liquids (bromoform: $\sim 2.89 \text{ g/cm}^3$ and diiodomethane: $\sim 3.32 \text{ g/cm}^3$) to concentrate apatite fraction.

Apatite mounts were prepared using the external detector method (EDM), which utilizes a Zeta calibration factor (Hurford & Green, 1983). This factor was determined by analyzing the Fish Canyon and Durango age standards at the Thermochronology Laboratory of the University of Padova, Italy. Crystals were selected using an Olympus SZX16 binocular microscope, mounted on double-sided tape, embedded in epoxy sheets, and polished to expose their internal surfaces. The exposed surfaces were then etched with a solution 5N of nitric acid HNO_3 during for 20 seconds at 20°C to etch fission tracks. The mountings were covered with mica detectors and irradiated in two separate campaigns (OSU28, OSU29B) at the Oregon State Radiation Center, USA, with nominal fluences of $1.40 \times 10^6 \text{ n/cm}^2$ and $1.22 \times 10^6 \text{ n/cm}^2$, respectively. The irradiations tube included Fish Canyon, Mount Dromedary, and Durango age standards, as well as CN5 dosimeter glasses. After irradiation, the mica detectors were detached, etched with 40% HF at 20°C for 45 minutes, and prepared for track counting.

Fission tracks were counted dry at 1250× magnification using an Olympus BX-51 optical microscope with an automated XY-stage system at the University of Padova. Ages were calculated using a zeta value of 356.04 ± 14.58 (zeta determined by analyst Sandoval Leidy C. for the CN5 dosimeter glass). Under the same system, etch pit diameters (D_{par}) were measured, and where possible confined fission tracks lengths were obtained. AFT ages are reported as central ages, calculated using the radial plot software developed by Vermeesch (2018). Detailed AFT results are summarized in Table 2.8 with associated radial plots shown in Fig. 6.2.

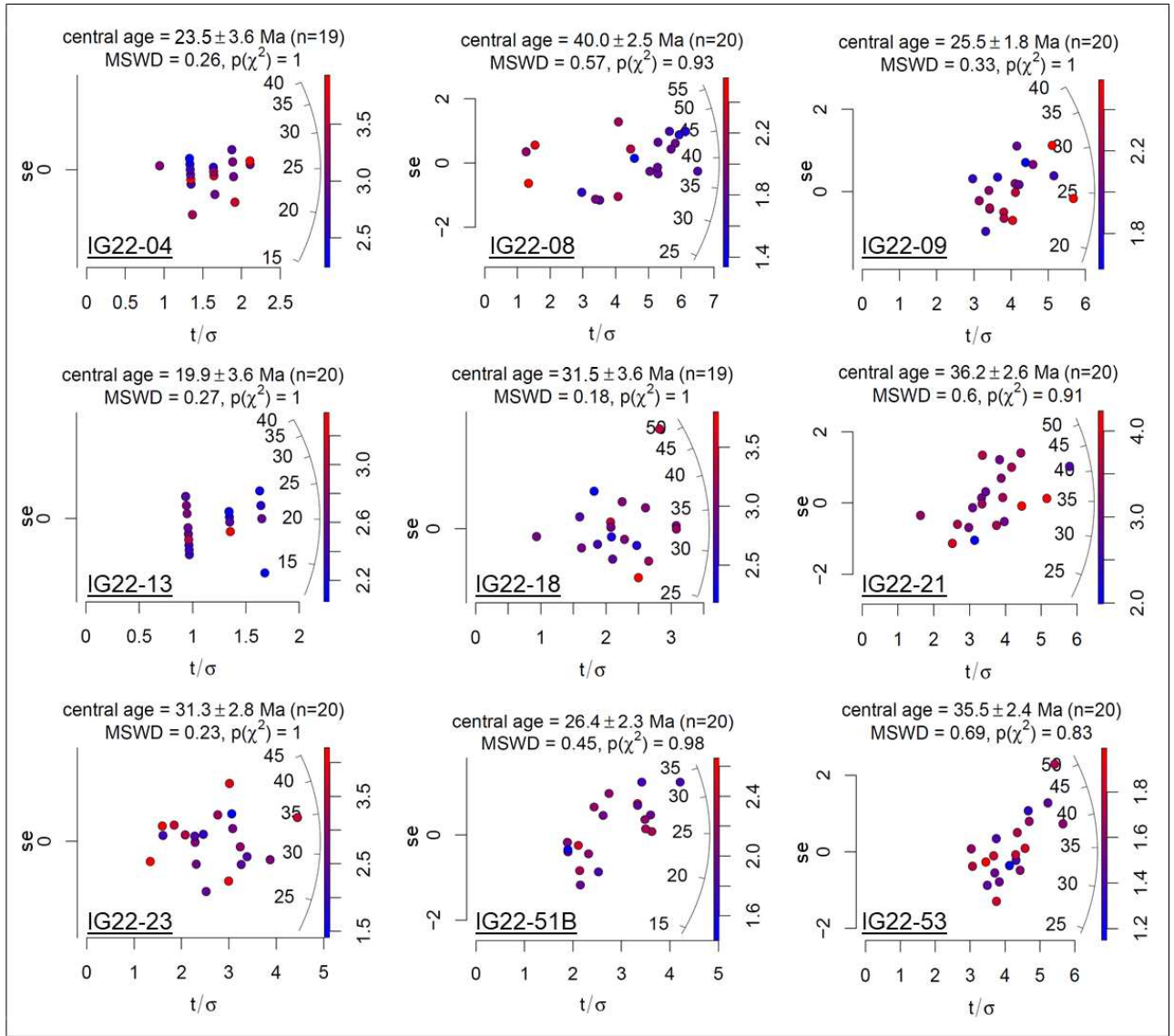


Figure 6.2 Radial plots of apatite fission track (AFT) data for selected samples from this study, illustrating single-grain age distributions. The bar on the right of each radial plot represents the Dpar in micrometers.

6.3.2 Thermal modeling

Thermal histories were reconstructed using the HeFTy software version 2.1.7 developed by Ketcham (2005), which employs a 'Frequentist' approach that uses formal statistical hypothesis tests to assess the goodness-of-fit between the input data and the thermal model predictions. We applied the annealing model of Ketcham et al. (2007), used etch pit diameters (Dpar) as kinetic parameters (Donelick et al., 2005), and aimed to minimize constraints in the model to avoid over-constraining.

To guide the thermal history modeling, we defined two primary constraint boxes for all samples. These constraints reduce model ambiguity by integrating independent geological and

geochronological evidence. The first constraint was based on U–Pb zircon crystallization ages from the same samples (Iglesias et al., 2025, submitted; Sandoval-Espinel et al., 2025, in press). The second constraint represents the present-day surface temperature, fixed at 20 ± 5 °C.

For samples from the ATB (samples IG22-53, IG22-07, IG22-08, and IG22-09), we applied an additional constraint derived from stratigraphic and structural evidence indicating that the massif was exposed at the surface during the Late Cretaceous (Albian–Maastrichtian) and was unconformably overlaid by the Puyango–Cazaderos and Cochurco formations (Jaillard et al., 1999; Riel et al., 2014).

For samples from the Western Cordillera (IG22-21 and IG22-23), we incorporated previously published AHe data from Spikings et al. (2010) into the thermal models to improve temporal constraints on the cooling histories. For sample IG22-21, AHe data from samples 99RS40 and 99RS41, located within ~5 km of our sampling site, were included. For sample IG22-23, AHe data from sample 99RS43, located within ~15 km, were incorporated. The inclusion of these nearby samples is justified by their position within the same structural block and their consistent lithological context, supporting the assumption that they experienced a similar thermal history.

Models were run until 500 good paths (defined as those that statistically fit the observed AFT ages and length distributions within the imposed constraints, $p\text{-value} \geq 0.05$, and GOF of ~ 1.0) were found. Samples IG22-15 and IG22-17 were excluded due to having too few grains (fewer than 5) and no measured confined track lengths.

Finally, to estimate exhumation rates, we used the Age2EDOT program (Brandon et al., 1998; Ehlers et al., 2005), which models cooling rates and calculates the apparent geothermal gradient (dT/dZ , °C/km) for a given thermochronometric system. As input parameters we assumed a surface temperature of 20 °C and a geothermal gradient of 35 °C/km, consistent with the elevated heat flow observed in the present-day southern Ecuadorian magmatic arc. Given the long-lived magmatic activity in this region and the absence of evidence for significant thermal regime changes, it is reasonable to infer that past geothermal gradient were comparable to current values, which typically exceed 30 °C/km (e.g., Angulo-Romero et al., 2023).

6.4 Results

6.4.1 AFT thermochronology

AFT ages range from 40.3 ± 2.5 Ma (sample IG22-08; Fig. 2) to 15.4 ± 1.9 Ma (sample IG22-02). In general, samples from the Celica–Lancones Arc and the Eastern Cordillera show a positive correlation between age and elevation (Figs. 6.3A and 6.3E), while samples from the ATB display a negative correlation. In the Western Cordillera, the three samples form a boomerang pattern with no clear age–elevation relationship. MTL range from $12.89 \mu\text{m}$ to $14.18 \mu\text{m}$ (Fig. 6.3B), and Dpar values range from $1.8 \mu\text{m}$ to $3.0 \mu\text{m}$ (Fig. 6.3C). All samples passed the $p(\chi^2)$ test, indicating the presence of a single population.

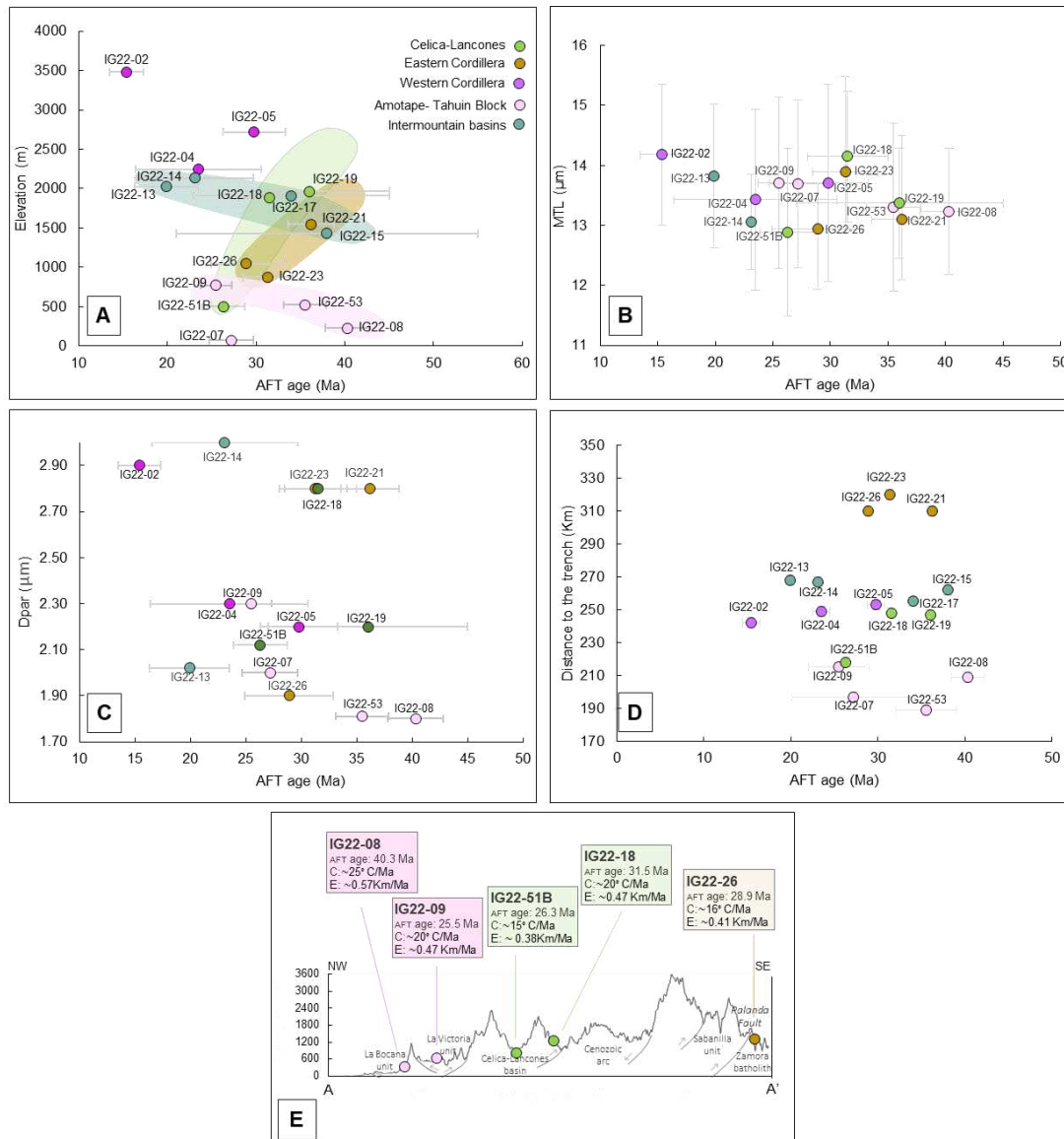


Figure 6.3 (A) Age-elevation relationships, (B) Mean track length (MTL) vs. AFT age, (C) Dpar vs. AFT age, and (D) Distance to the trench vs. age (E) Topographic profiles illustrating sample locations. The locations of the profiles are shown in Figure 6.1C. C: cooling rate. E: exhumation rate.

Four AFT ages were obtained from the ATB (Fig. 1B), including three samples from the El Oro Complex. The Triassic granite (IG22-53) yielded an AFT age of 35.5 ± 2.4 Ma and a MTL of 13.3 ± 1.4 μm . The granodiorite (IG22-08) gave an age of 40.3 ± 2.5 Ma and an MTL of 13.2 ± 1.1 μm , whereas the granite (IG22-09) yielded 25.5 ± 1.8 Ma and an MTL of 13.7 ± 1.4 μm . The northernmost sample (IG22-07), from the Biron Complex, yielded an age of 27.2 ± 2.5 Ma with an MTL of 13.7 ± 1.7 μm . These results indicate protracted cooling of the ATB from the late Eocene to early Miocene.

Three plutonic samples from the Western Cordillera were analyzed (Fig. 6.1B). The Oligocene quartz-diorite (IG22-04) gave an AFT age of 23.5 ± 3.6 Ma and an MTL of 13.7 ± 1.6 μm , while sample IG22-05 yielded an AFT age of 29.8 ± 7.1 Ma with an MTL of 13.4 ± 1.5 μm . The Miocene granodiorite IG22-02 yielded an AFT age of 15.4 ± 1.9 Ma and an MTL of 14.2 ± 1.2 μm ; this age closely matches its zircon U–Pb crystallization age of 13.3 ± 0.2 Ma, indicating rapid post-magmatic cooling.

Three Late Cretaceous samples from the Celica–Lancones Arc (gabbro IG22-18, granodiorite IG22-19, diorite IG22-51B; Fig. 6.1B) yielded AFT ages ranging from 35.9 ± 9.7 Ma to 26.3 ± 2.3 Ma, with MTL values between 13.3 ± 1.4 μm and 14.1 ± 1.5 μm . Additionally, sample IG22-19, despite having 13 counted apatite grains (typically sufficient for igneous rocks), exhibits an unusually large age uncertainty, likely related to variable track preservation, heterogeneous uranium distribution, or analytical limitations.

Four granitoids samples from the Intermontane Basins (IG22-13, IG22-14, IG22-15, IG22-17; Fig. 6.1B) yielded AFT ages between 19.9 ± 3.6 Ma and 38 ± 11 Ma; these overlap with their respective zircon U–Pb crystallization ages (Iglesias et al. 2025, submitted), suggesting mainly post-magmatic cooling.

Three granodiorite samples from the Eastern Cordillera yielded Oligocene to Eocene AFT ages. Samples IG22-26 and IG22-23, located east of the Palanda Fault (Fig. 6.1), yielded similar Oligocene ages of 28.9 ± 4.0 Ma and 31.3 ± 2.8 Ma, respectively, whereas sample IG22-21, west of the fault, present an older Eocene age of 36.2 ± 2.6 Ma. These differences may reflect differential exhumation across the Palanda Fault during the late Eocene to Oligocene

6.4.2 Thermal history models

Thermal modeling reveals distinct cooling events across different crustal domains (Fig. 6.4), which are interpreted as exhumation episodes, as cooling in orogenic settings commonly reflects the uplift and erosion of deeper, hotter rocks toward the surface.

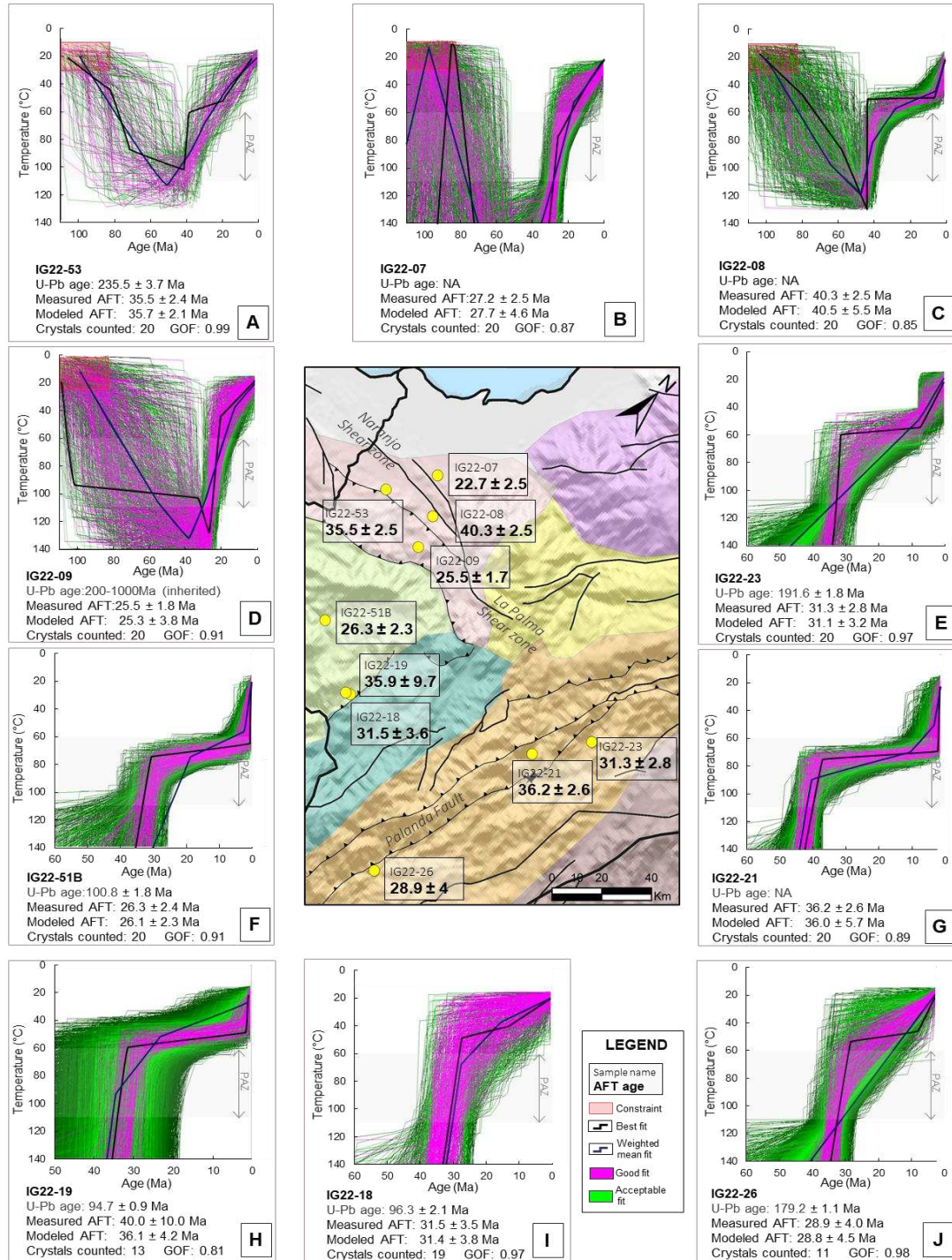


Figure 6.4 Map showing the location of the samples analyzed in this study. For each sample that records exhumation-related cooling, the corresponding time-temperature (T-t) histories modeled using HeFTy (Ketcham, 2005) is shown. Model constraints are included (see text for details); however, U-Pb crystallization age constraints are not displayed, as they fall outside the plotted time range. PAZ: partial annealing zone.

A) Amotape-Tahuín Block (ATB)

Samples IG22-53 (Fig. 6.3A) and IG22-08 (Fig. 6.4C) record similar thermal histories. After acting as sources for the Albian–Maastrichtian sedimentary formations (Jaillard et al., 1999; Riel et al., 2014), both samples experienced reheating, followed by cooling between ~45 - 40 Ma. Although some good-fit paths show thermal quiescence between ~60 and 40 Ma. The best-fit models indicate rapid 42–40 Ma cooling at ~35 °C/Ma, corresponding to exhumation of ~0.7 km/Ma considering a geothermal gradient of 35 °C/km.

In contrast, the northernmost (IG22-07; Fig. 6.4B) and southernmost (IG22-09; Fig. 6.4D) samples record reheating followed by later cooling, at ~33–25 Ma. Their best-fit models yield rates of ~30–35 °C/Ma (equivalent to ~0.6 -0.7 km/Ma of exhumation). Mean track lengths >13 μ m (Fig. 6.3B, Table 2) across the ATB indicate short residence times within the partial annealing zone, consistent with relatively rapid exhumation.

B) Celica-Lancones arc

Samples from this block (IG22-51B, IG22-19, and IG22-18; Figs. 6.4F, 6.4H, and 6.4I, respectively) show cooling primarily between ~40 and 29 Ma, as indicated by the majority of good-fit paths, with only a few paths suggesting cooling in the ~50–40 Ma interval. Best-fit models for all samples indicate cooling initiated at ~35–33 Ma, at rates of ~15–30 °C/Ma, equivalent to exhumation of ~0.4–0.6 km/Ma.

C) Eastern Cordillera

Samples IG22-23 and IG22-26, located east of the Palanda Fault (Figs. 6.1 and 6.4), exhibit broadly similar thermal histories. Sample IG22-23 (Fig. 6.4E) shows cooling mainly between ~38 and 32 Ma, as indicated by most of the good-fit paths. Sample IG22-26 (Fig. 6.4J) records a similar primary cooling interval, although a few paths extend to around 42 Ma. The best-fit models for these samples indicate cooling between ~34 and 32 Ma, with rates of about 16–20 °C/Ma, corresponding to exhumation rates of 0.4–0.5 km/Ma. In contrast, sample IG22-21, located west of the Palanda Fault, records an older cooling episode. All good-fit paths (Fig. 6.4G) indicate cooling between ca. 45 and 40 Ma, while the best-fit model suggests cooling at around 42 Ma, with a rate of about 20 °C/Ma, corresponding to an exhumation rate of ~0.5 km/Ma. A potential change in cooling rate is suggested by best-fit paths of all studied samples at temperatures lower than ~60 °C; however, this occurs out of the sensitivity range of the method employed in this study and cannot be evaluated further.

6.5 Discussion

6.5.1 Cooling events within the crustal blocks

The interpretation of regional exhumation patterns must be considered in light of sampling density. Although each major crustal block is represented, only 1–2 samples per block yielded suitable apatite concentrates, as many collected rocks contained apatites that were highly fractured, inclusion-rich, or otherwise unsuitable for reliable track counting. This limits high-resolution assessment of intra-block variability; however, the dataset still constitutes one of the most comprehensive AFT compilations for southern Ecuador and provides a robust first-order framework for reconstructing the region's cooling history. Our thermochronological dataset indicates broadly continuous and asynchronous exhumation across southern Ecuador (Fig. 6.5A), with cooling occurring in partially overlapping intervals across different crustal blocks (Figs. 6.4 and 6.5). Within this long-term exhumation history, three main cooling clusters can be recognized: (i) a middle- late Eocene interval (~45–38 Ma; Fig. 6.5D) characterized by rapid cooling of ~20–35 °C/Ma; (ii) a late Eocene– early Oligocene interval (~40–30 Ma; Fig. 6.5E) with cooling rates of ~15–30 °C/Ma; and (iii) Oligocene interval (~33–25 Ma; Fig. 6.5F), during which cooling rates increased again to ~30–35 °C/Ma. Beyond the exhumation-related cooling patterns, rapid post-magmatic cooling is also recorded in some structural blocks (Fig. 6.5A).

6.5.1.1 Eastern Cordillera

Cooling in the southern Eastern Cordillera was evaluated in two units: the Jurassic Zamora Batholith and the Triassic Sabanilla Unit. The Zamora Batholith, located east of the Palanda Fault (footwall; Figs. 6.1B and 6.4), records a cooling interval between ~38 and 32 Ma, with best-fit thermal models for samples IG22-23 and IG22-26 indicating rates of ~16–23 °C/Ma (Figs. 6.4E and 6.4J). Within the same structural block, Spikings et al. (2001) reported an AFT age of 34 ± 3.3 Ma and MTL of 12.6 ± 0.5 μm for sample 99RS43, which is consistent with our results, although they inferred slower cooling rates (10°C/Ma). Farther east within the same batholith, along the Sub-Andean zone, the same authors documented older AFT ages between 61–59 Ma, suggesting an earlier phase of exhumation in that region. To the west of the Palanda Fault (hanging wall), the Sabanilla Unit experienced cooling between 45 and 40 Ma at a rate of ~20°C/Ma (sample IG22-21; Fig. 6.4G). This event is consistent with the 45–30 Ma cooling phase at ~20°C/Ma reported by Spikings et al. (2010) for the same unit. These results reveal distinct cooling histories across the Palanda fault, indicating that differential exhumation within the Eastern Cordillera was likely controlled by reactivated faults systems inherited

from earlier tectonic phases. The findings also suggest that exhumation in the Eastern Cordillera began as early as the Paleocene and continued through the late Eocene–early Oligocene.

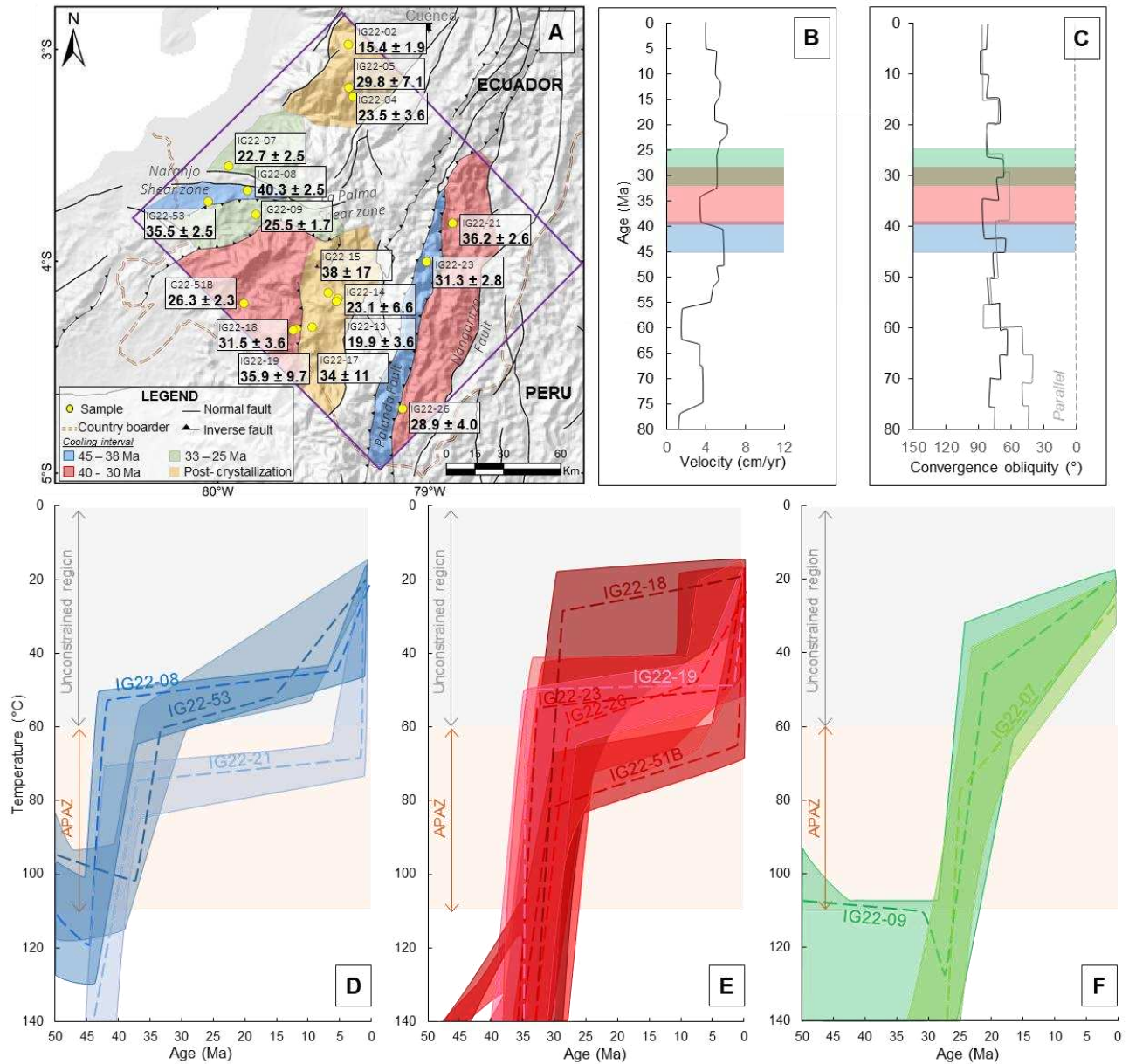


Figure 6.5 (A) Map of the study area showing crustal blocks color-coded according to their corresponding cooling intervals. (B) Normal convergence rates over time for the Nazca Plate (formerly Farallón) and the South American Plate, from Maloney et al. (2023). (C) Convergence obliquity versus age, after Jaillard (2025). (D–F) Summary of all good T–t paths identified in this study (shaded areas); dashed lines represent best-fit paths for individual samples. Shaded areas correspond to distinct cooling intervals: blue (45–38 Ma), red (40–30 Ma), and green (33–25 Ma). APAZ = apatite partial annealing zone.

6.5.1.2 Intermountain basins

In the Intermontane region of southern Ecuador, igneous rocks represented by samples IG22-13 (AFT age: 19.9 ± 3.6 Ma; U-Pb age: 18.5 ± 0.2 Ma), IG22-14 (AFT age: 23.1 ± 6.6 Ma; U-Pb age: 21.7 ± 0.4 Ma), and IG22-15 (AFT age: 38.0 ± 17 Ma; U-Pb age: 36.7 ± 0.4 Ma) exhibit AFT and U-Pb ages

that are broadly similar. This close correspondence indicates rapid post-crystallization cooling (Figs. 6.5A and 6.6), suggesting that these granitoids were emplaced at shallow crustal levels, likely near the PAZ. Therefore, the AFT ages likely record thermal relaxation following magmatic emplacement rather than significant exhumation (e.g., Malusà & Fitzgerald, 2019). However, coeval tectonically controlled exhumation cannot be completely ruled out, particularly in the context of ongoing convergence (Jaillard, 2025; and references therein), significant crustal thickening and exhumation recorded along the Northern Andes during the Miocene (e.g., Bermúdez et al., 2011; Horton et al., 2020; Margirier et al., 2023; Mora et al., 2010; Sandoval-Espinel et al., 2024; Spikings et al., 2000; Spikings et al., 2001). Throughout this period, some segments of the Northern Andes experienced the complete emergence of topographic features, leading to the establishment of the Andean belt as an effective orographic barrier (e.g., Horton et al., 2020; Witt et al., 2017).

6.5.1.3 Celica-Lancones arc

The Cenomanian rocks of the Celica-Lancones arc, located between the Intermontane basins to the east and the ATB to the west (Fig. 6.1), underwent synchronous cooling and exhumation between ~40 and 29 Ma, at estimated rates of ~15–20 °C/Ma. Although previous work in this crustal block is limited, a single detrital thermochronology dataset (Espurt et al., 2018; sample 7) suggests a possible uplift phase around 30 Ma. Our results support this timing and provide the first direct thermochronological evidence of early Oligocene exhumation in the region derived from igneous bedrock.

6.5.1.4 Western Cordillera

Samples from the Western Cordillera yielded AFT ages of 29.8 Ma (IG22-05) and 23.5 Ma (IG22-04). While no U–Pb crystallization age is available for IG22-05 due to the presence of only inherited zircon components (~1 Ga), sample IG22-04 provides a more complete thermal history. Its zircon U–Pb age (29.7 ± 2 Ma) closely matches its AFT age (23.5 ± 3.6 Ma), suggesting rapid post-magmatic cooling, with emplacement and cooling occurring nearly simultaneously. Since both samples belong to the same crustal block, they likely record a shared tectono-thermal history, most plausibly reflecting post-magmatic thermal relaxation following magmatism.

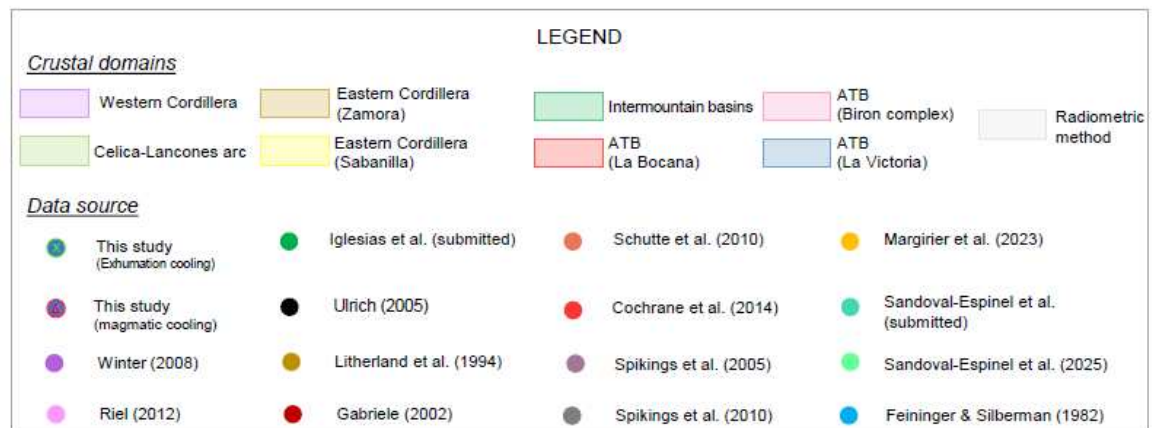
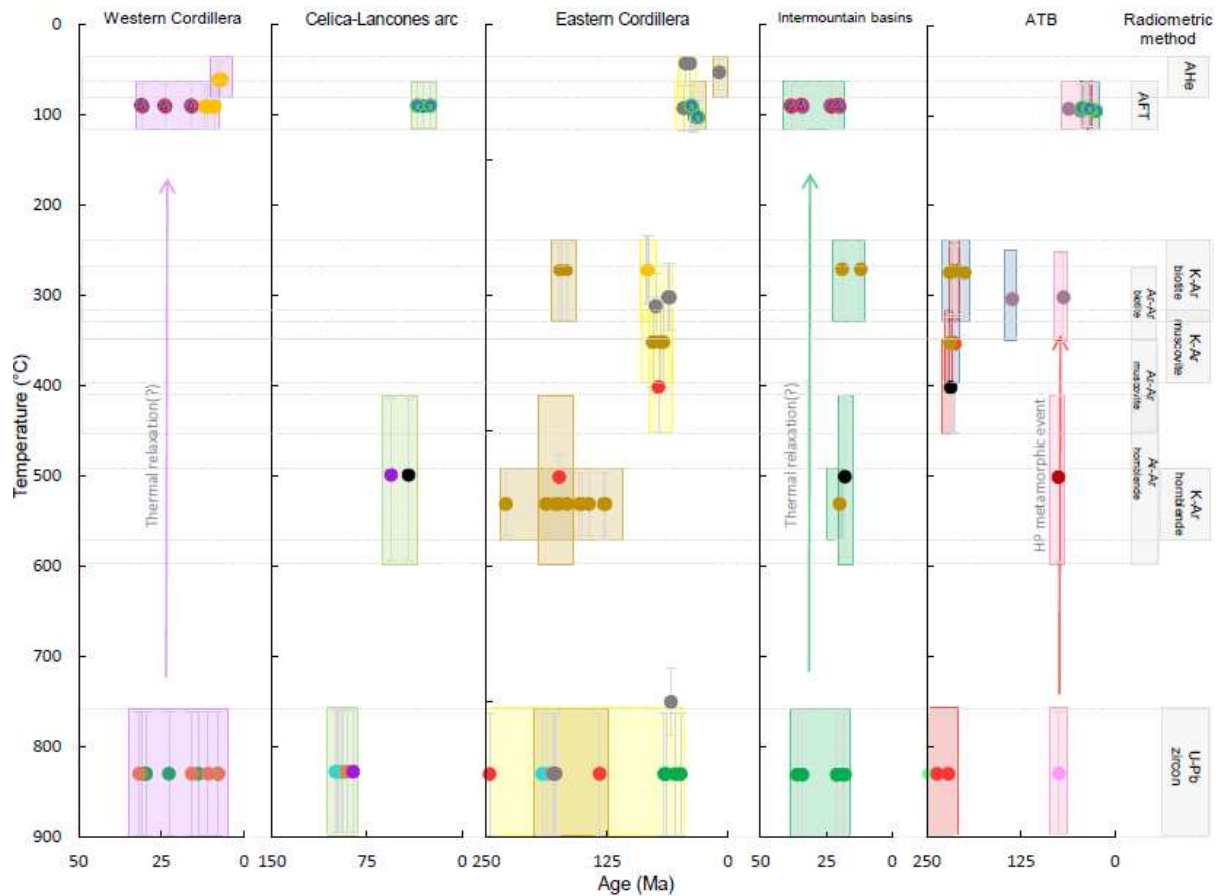


Figure 6.6 Cooling temperature (°C) versus age (Ma) plots showing geochronological and thermochronological ages from in situ rocks across different tectonostratigraphic units of southern Ecuador. Ages are from Cochrane et al. (2014), Feininger & Silberman (1982), Gabriele (2002), Iglesias et al. (2025, submitted), Litherland et al. (1994), Margirier et al. (2023), Pratt et al. (2005), Riel et al. (2012, 2014), Sandoval-Espinel et al. (2025, submitted), Schutte et al. (2010), Spikings et al. (2001, 2005, 2010), Winter (2008), and Ulrich (2005).

This interpretation aligns well with results from Margirier et al. (2023), who studied the same crustal block slightly north of our study area (~3°S) and documented an early–middle Miocene post-magmatic cooling phase attributed to thermal relaxation after intrusion. Our results reinforce the idea that AFT thermochronology in this region primarily records cooling related to the final stages of magmatism rather than tectonic exhumation (Fig 6.5A).

6.5.1.5 Amotape-Tahuin Block

The ATB is unconformably overlain by the Puyanco-Cazaderos and Cochurco series, indicating that the massif was at the surface during the Aptian to Coniacian or Santonian (Jaillard et al., 1999; Riel et al., 2014). Heating between ~80 and 45 Ma (Fig. 6.3) likely resulted from burial beneath Paleocene rocks, including the Gonzanama and Sacapalca formations (Hungerbühler et al., 2002; Jaillard et al., 1999), as there is no evidence of an increased geothermal gradient, such as volcanism, within the ATB during this period (Fig. 6.6). These findings align with a post-Late Cretaceous burial-related reheating phase previously documented by Espurt et al. (2018) in a Triassic granite of northern Peru. Following this, the ATB underwent heterogeneous cooling and exhumation.

The northernmost unit of the ATB, located north of the La Palma shear zone (Fig. 6.1B) experienced cooling between ~33 and 25 Ma at rates of ~30°C/Ma. Spikings et al. (2005) reported an AFT and ZFT age of 61 and 68 Ma, respectively, in the same block, while Gabriele (2002) obtained an age $^{40}\text{Ar}/^{39}\text{Ar}$ hornblende age of 75 ± 0.5 Ma. Documenting amphibolitized metagabbroic bodies in the area, Riel et al. (2014) attributed the ~75-73 Ma ages to high-temperature deformation. Their interpretation suggests either an accretion event or a poorly understood, localized tectonothermal event that might have partially overprinted pre-existing thermal and structural signatures. However, the AFT age of 61 ± 5 Ma and MTL of 12.4 ± 0.2 reported by Spikings et al. (2005) may not be directly linked to this slightly older high-temperature event. In fact, Spikings et al. (2005) noted that their AFT ages from the ATB generally do not correspond to specific, well-defined cooling events. Their ages were interpreted as reflecting partially annealed fission tracks, suggesting that the samples experienced incomplete thermal resetting. This could explain the older apparent age despite the close spatial proximity to our sample. In contrast, the sample analyzed in this study likely records a more recent and distinct cooling episode, as indicated by its higher Chi-square (χ^2) probability value and longer MTL (13.5 μm). The observed difference in AFT ages may therefore reflect localized thermal processes, such as hydrothermal activity, or structural influences, such as faulting. Additional data are required to better constrain these interpretations.

The E-W trending block of the ATB, located just south of the NZS, experienced a rapid cooling event between 45 and 40 Ma at rates of ~35°C/Ma. This event is recorded in both the La Bocana Unit and the Triassic La Florida Pluton. Within the same block, Spikings et al. (2005) reported an AFT age of 38.8 ± 2.7 and MTL of 13.63 ± 0.19 μm , documenting the onset of cooling at ~43–39 Ma but at significantly lower rates of ~5.5–4.5°C/Ma. The difference in cooling rates, despite similar AFT ages, is likely to be related to the constraints and assumptions used in the thermal modelling. Further south,

the block comprising the La Victoria Unit (Fig. 6.1B) underwent cooling between ~33 and 25 Ma at rates of 30–35°C/Ma.

6.5.2 Cooling history of southern Ecuador and its relationship to Andean tectonic events

The arrival of the Pallantanga Terrane at the northwestern margin of South America around 75–73 Ma (Cediel, 2019; Jaillard, 2022; Luzieux et al., 2006; Vallejo et al., 2019; Witt et al., 2023) marked the onset of cooling and exhumation in several regions along the northern Andes (i.e., Cediel & Shaw, 2019; Luzieux et al., 2006; Spikings et al., 2010; Witt et al., 2017). This event is thought to be recorded in southern Ecuador along the eastern margin of the Eastern Cordillera and the Sub-Andean Zone, within faulted blocks of the Zamora Batholith (Spikings et al., 2001; 2010). It also coincides with the deposition of the continental Tena Formation (72–57 Ma; Balkwill et al., 1995) in the Amazon foreland basin, which is interpreted as the first rocks sourced from the Eastern Cordillera, indicating that the Cordillera was uplifted during this period (Ruiz et al., 2004; Spikings et al., 2010). Prior to the Maastrichtian, rocks from the Amazon basin were derived from the Brazilian Shield (Martin-Gombojav & Winkler, 2008; Ruiz et al., 2004; Vallejo et al., 2021). However, our dataset does not record a Maastrichtian cooling signal. This absence may reflect the limited sampling of more easterly structural blocks, where such cooling has been suggested, or it may simply indicate that southern Ecuador does not preserve clear evidence of this northern accretionary event.

A reorientation of the Farallon plate convergence vector from NNE or NE to NE or ENE (~45–40 Ma; Fig. 6.5C; Jaillard, 2025; and references therein), significantly increased compression in the upper plate. This tectonic reorganization might have influenced the middle-late Eocene cooling event (~45–38 Ma), that we captured particularly on the western flank of the Eastern Cordillera (Sabanilla unit) and the core of the ATB (Fig. 6.7A). This event culminated with the deposition of unconformable conglomerates and volcanic rocks of late Eocene age, such as the Quingeo Formation (42–34 Ma; Hungerbühler et al., 2002; Jaillard, 2025; and references therein).

Subsequently, around 38 Ma, a slight increase in convergence rates (Fig. 6.5B; Maloney et al., 2013) likely contributed to the exhumation of the Celica Arc and the crustal block of Zamora Batholith bounded by the Nangaritza and Palanda faults (Fig. 6.7B). This event is evidenced by significant paleogeographic changes along the Ecuadorian margin and the northern Peruvian Andes, including widespread erosion and a shift from marine to terrestrial depositional environments (Espurt et al., 2018; Witt et al., 2019). This period has been interpreted as a major compressional event (Jaillard et al., 1995; Naeser et al., 1991). In southern Ecuador, late Eocene to early Oligocene compression has

been specifically linked to a shift in lag-times (difference between the AFT and the sedimentation age, see Malusà & Fitzgerald, 2020) recorded in the Tiyuyaku Formation of the Amazon foreland basin. Around ~36 Ma, lag times changed from relatively short values between ~50–36 Ma to progressively increasing lag times from ~36–30 Ma (see Ruiz, 2002; Spikings et al., 2010). This transition has been interpreted as reflecting a change in sediment provenance within the Eastern Cordillera. However, the results of this study indicate a decrease in exhumation rates, from higher cooling rates of ~20–35 °C/Ma during the middle- late Eocene (~45–38 Ma) to lower rates of ~15–30 °C/Ma during the late Eocene to early Oligocene (~40–30 Ma). Therefore, the observed increase in lag times may reflect slower exhumation rather than a shift in the sediment source region.

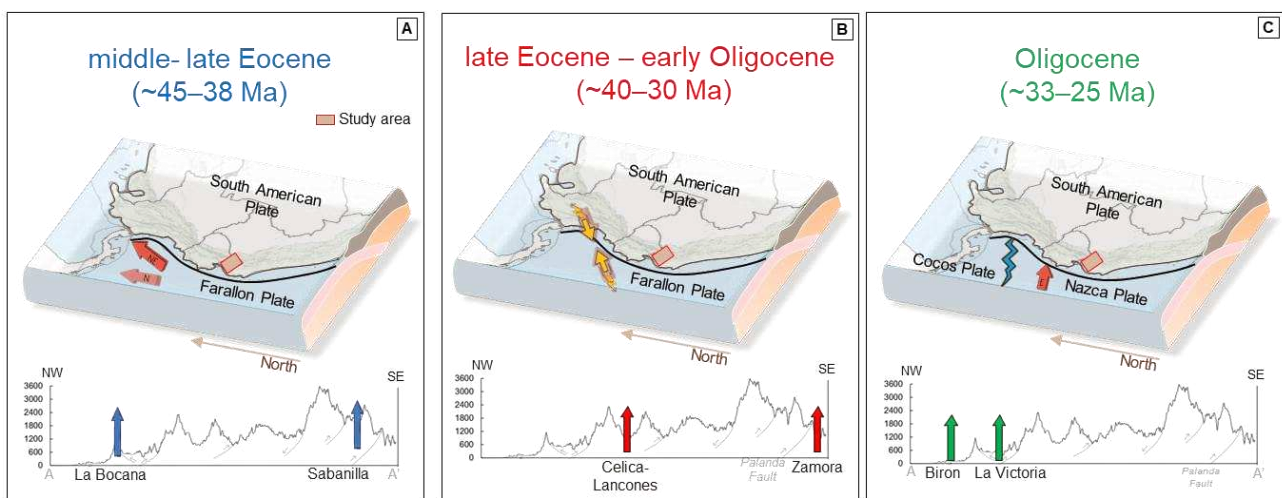


Figure 6.7 Cartoon summary of major cooling intervals in southern Ecuador, showing the inferred causes for each interval and a profile of the study area (see Figure 1 for location) indicating blocks that experienced cooling (upward-pointing arrows). (A) middle Eocene event illustrating a change in subduction vector direction. (B) late Eocene–Oligocene event depicting an increase in subduction rate. (C) late Oligocene–early Miocene event showing a renewed change in subduction vector direction.

A tectonic plate rearrangement about 27–25 Ma, driven by the fragmentation of the Farallon Plate (Jaillard, 2025; Maloney et al., 2013) resulted in a shift in the subduction direction from ESE to E (Gordon & Jurdy, 1986; Pilger 1984). This reconfiguration broadly coincides with the Oligocene (~33–25 Ma) exhumation event, which led to the exhumation of the peripheral blocks of the ATB in southern Ecuador (Fig. 6.7C). Furthermore, this event broadly coincides with a major arc magmatic flare-up event in Ecuador that occurred during the late Oligocene to early Miocene and was characterized by widespread ignimbrite eruptions and pluton emplacement (Schütte et al., 2010). Crustal thickening associated with magmatic addition can lead to surface uplift through isostatic compensation (see Farner & Lee, 2017). Additionally, magmatism can drive uplift through several mechanisms, including diapirism and crustal inflation associated with pluton emplacement, as well

as decompression melting and magma generation triggered by basal lithospheric removal. For example, Perkins et al. (2016) documented crustal thickening and uplift in the Central Andes caused by the accumulation of buoyant magmatic bodies. While, Studies by Coldwell et al. (2011) and Garzione et al. (2006) in central Peru and Bolivia, respectively, have shown that removal of the lower continental crust and mantle lithosphere (via subduction erosion and delamination) induced asthenospheric upwelling, decompression melting, significant magmatic activity and ultimately surface uplift. The relative contributions of magmatism and tectonic processes to uplift in the southern Ecuadorian Andes remain poorly constrained, largely due to uncertainties in the volume and spatial distribution of the intrusive and volcanic material produced during the flare-up event.

6.5.3 Heterogeneous uplift and topographic development of southern Ecuador

Results from this study indicate that exhumation in southern Ecuador was already active by at least the middle Eocene. Although our thermochronological dataset (AFT ages) primarily constrains cooling rather than absolute surface uplift, the integration of stratigraphic evidence supports a heterogeneous pattern of topographic development across the region.

By the middle Miocene, large shallow-marine embayment (e.g., Loja) developed under extensional conditions and experienced marine incursions from the subsiding forearc (Progreso Basin; Steinmann et al., 1999). These processes led to the accumulation of coastal to shallow-marine facies between ~15 and 10 Ma within the southern intermontane basins. Stratigraphic evidence suggests that these coastal-marine systems were mainly supplied with detritus derived from the Eastern Cordillera and surrounding older volcanic formations (Hungerbühler et al., 2002; Steinmann et al., 1999), probably including material from the Celica–Lancones Basin, which is dominated by continental and volcanoclastic successions. The absence of middle Miocene marine deposits in the Celica–Lancones Basin (Jaillard et al., 1999) further supports the interpretation that both the Eastern Cordillera and the Celica–Lancones domain were already partially uplifted by this time. These structural highs likely formed the topographic margins of the Miocene Loja embayment.

A regionally extensive angular unconformity dated between ~10 and 8 Ma across the intermontane basins (Hungerbühler et al., 2002) marks the onset of basin inversion and continental sedimentation, reflecting a major phase of uplift (Hungerbühler et al., 2002; Steinmann et al., 1999). Exhumation events younger than ~10 Ma enhanced topographic growth, through renewed compressional deformation, crustal shortening and erosion (Steinmann et al., 1999). These processes uplifted the 15–10 Ma shallow-marine deposits to their present elevations of ~2700 m. This evolution is consistent

with regional evidence indicating that the main phase of Andean topographic development occurred during the Miocene (e.g., Anderson et al., 2016; Echeverri et al., 2015; Giambiagi et al., 2022; Howlett et al., 2025; Mosolf et al., 2011). A more quantitative assessment of the uplift magnitude, rates, and spatial variability in this Andean segment requires future work. This work should integrate apatite (U–Th)/He thermochronology, cosmogenic-nuclide analyses, and both thermokinematic and landscape-evolution modeling.

6.6 Conclusions

Apatite fission-track thermal modeling from southern Ecuador reveals a pattern of continuous cooling from the middle Eocene to the present, punctuated by three distinguishable cooling–exhumation intervals: ~45–38 Ma (middle Eocene), ~40–30 Ma (late Eocene to Oligocene), and ~33–25 Ma (late Oligocene to early Miocene). Two of these exhumation phases (~45–38 Ma and ~33–25 Ma) broadly coincide with documented changes in the direction of Nazca–South America plate convergence, whereas the ~40–30 Ma interval overlaps with an increase in convergence rate (Maloney et al., 2013).

Exhumation was spatially heterogeneous and varied across distinct crustal blocks. In the Eastern Cordillera, the Jurassic Zamora Batholith cooled during the late Eocene–Oligocene, whereas the adjacent Triassic Sabanilla Unit (structurally separated by the Palanda Fault) experienced earlier, middle-Eocene exhumation. A similar pattern characterizes the Amotape–Tahuin Block, where the core cooled during the middle Eocene, followed by younger late Oligocene–early Miocene exhumation in the fault-bounded peripheral domains. These observations indicate that fault activity modulated exhumation, accommodating deformation during regional tectonic reorganization. Finally, the temporal overlap between the late Oligocene–early Miocene exhumation phase and a proposed magmatic flare-up (Schütte et al., 2010) suggests that magmatism may have contributed to exhumation. Together, these results indicate a dual influence of external plate-convergence dynamics and internal processes (magmatism and fault-driven block accommodation) in shaping the exhumation history of this segment of the Andes.

Declaration of Competing Interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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CHAPTER 7: THERMAL AND TECTONIC EVOLUTION OF THE SOUTHERN ECUADORIAN ANDES: NUMERICAL MODELING OF POST-MAGMATIC, AND EROSIONAL COOLING

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Declaration statement

The manuscript is in preparation; the journal has not yet been chosen.

CRedit authorship contribution statement

Sandoval Espinel, L.C.: Conceptualization, Investigation, Visualization, Data curation, Methodology, Writing – original draft. Bermudez, M.: Funding, Supervision, Writing – review & editing. Zattin, M.: Conceptualization, Supervision, Funding, Writing – review & editing. Witt, C.: Conceptualization, Supervision, Funding, Writing – review & editing.

ABSTRACT

The southernmost Andes of Ecuador record a complex thermo-tectonic history, resulting from changing subduction dynamics and internal crustal reorganization; however, uncertainties remain regarding the timing and rates of cooling events and their relationship with coeval magmatism. We applied 3D thermokinematic modeling (Pecube) and 4DTherm to existing bedrock thermochronology data to reconstruct the long-term thermal and exhumation history of two key regions: the Real Cordillera (Zabanilla Unit and Zamora Batholith) and the Amotape–Tahuin Block (La Bocana and La Victoria units). Additionally, 4DTherm modeling was used to distinguish exhumation-driven from magmatic to post-magmatic cooling along the southwestern Cordillera and intermontane basins. Our results reveal three main cooling pulses in the Southern Real Cordillera (~172 Ma at –1.63 km/Ma, ~63 Ma at 0.18 km/Ma, and ~9 Ma at 0.23 km/Ma), whereas the Amotape–Tahuin Block experienced multi-stage cooling with distinct phases in the La Bocana (~244 Ma at –0.62 km/Ma, ~95 Ma at 1.7 km/Ma, ~67 Ma at 1.2 km/Ma, and ~36 Ma at 0.12 km/Ma) and La Victoria (~235 Ma at –1.9 km/Ma, ~64 Ma at 0.34 km/Ma, and ~24 Ma at 0.26 km/Ma) units. In the Inter-Andean Valley and southwestern Cordillera, post-magmatic cooling rates were much higher than exhumation-related

cooling, consistent with shallow emplacement of magmatic bodies. Moreover, the increase in exhumation rates from the Late Eocene to the Late Oligocene–Early Miocene marks the possible onset of exhumation in the Inter-Andean Valley.

7.1 Introduction

The Southern Ecuadorian Andes has experienced a complex geodynamic evolution marked by alternating phases of extension and compression (e.g., George et al., 2021; Hungerbühler et al., 2002; Riel et al., 2014; Witt et al., 2023), magmatic activity (Sandoval-Espinel et al., 2025; Schütte et al., 2010; Sundell et al., 2022; Winter, 2008; Witt et al., 2017), and a progressive shift in subduction dynamics from highly oblique to more orthogonal convergence between the Nazca Plate (formerly Farallon Plate) and the South American Plate (Jaillard, 2025; Pardo-Casas & Molnar, 1987; Pilger 1984). This intricate tectonic regime has produced a complex and spatially variable exhumation history, as evidenced by previous studies (This study, chapter 6; Spikings et al., 2005, 2010), which suggest that exhumation was largely governed by major fault systems, leading to asynchronous cooling histories and variable exhumation rates across the region.

Despite general agreement on fault-controlled exhumation, discrepancies remain in the reported cooling histories. For example, in the Southern segment of the Cordillera Real (SCR), Spikings et al. (2005) proposed that the Zamora Batholith began cooling around 75 Ma and underwent slow cooling between 45 and 35 Ma at rates of approximately 10°C/Ma (corresponding to ~0.24 km/Ma, calculated using the AGE2EDOT code; Brandon et al., 1998; Ehlers et al., 2005). In contrast, we (Chapter 6) documented a younger and more rapid cooling event between ~34–31 Ma at rates of ~16–20°C/Ma, resulting in exhumation rates of ~0.4 to 0.5 km/Ma.

Similarly, for the Amotape-Tahuín Block (ATB; Fig. 7.1), we reported rapid cooling of the La Bocana Unit between 42 and 38 Ma, with estimated rates of approximately 25°C/Ma. In contrast, Spikings et al. (2005) documented significantly slower cooling rates for the same unit, ranging between ~4.5 and 5.5°C/Ma (~0.12 km/Ma). Further south, in the La Victoria Unit, we reported an AFT age of 25 Ma, whereas Spikings et al. (2005) obtained an older AFT age of ~45 Ma from the adjacent Triassic Marcabeli Pluton.

The apparent diachronism in timing, cooling, and exhumation rates within individual crustal blocks highlights the complexity and uncertainty of the exhumation history in southern Ecuador. Resolving these discrepancies are essential to developing a more consistent regional framework of exhumation in the southernmost part of the Northern Andes. Comparisons of 3D thermo-kinematic (Pecube) and 2D inverse thermal (4DTherm) models for intrusions is a promising approach to reconstruct from

magmatic, post-magmatic to erosional cooling of igneous intrusions directly from multiple thermochronometers. Unlike traditional inverse thermal models, such as HeFTy (Ketcham, 2005) and QTQt (Gallagher, 2012), allow us to reconstruct time-temperature paths at discrete sample locations without accounting for spatial variability or distinguishing between magmatic and exhumation-related cooling, tools like Pecube (Braun, 2003; Braun et al., 2006; Braun et al., 2012) and 4DTherm (Fu et al., 2010) offer more comprehensive solutions.

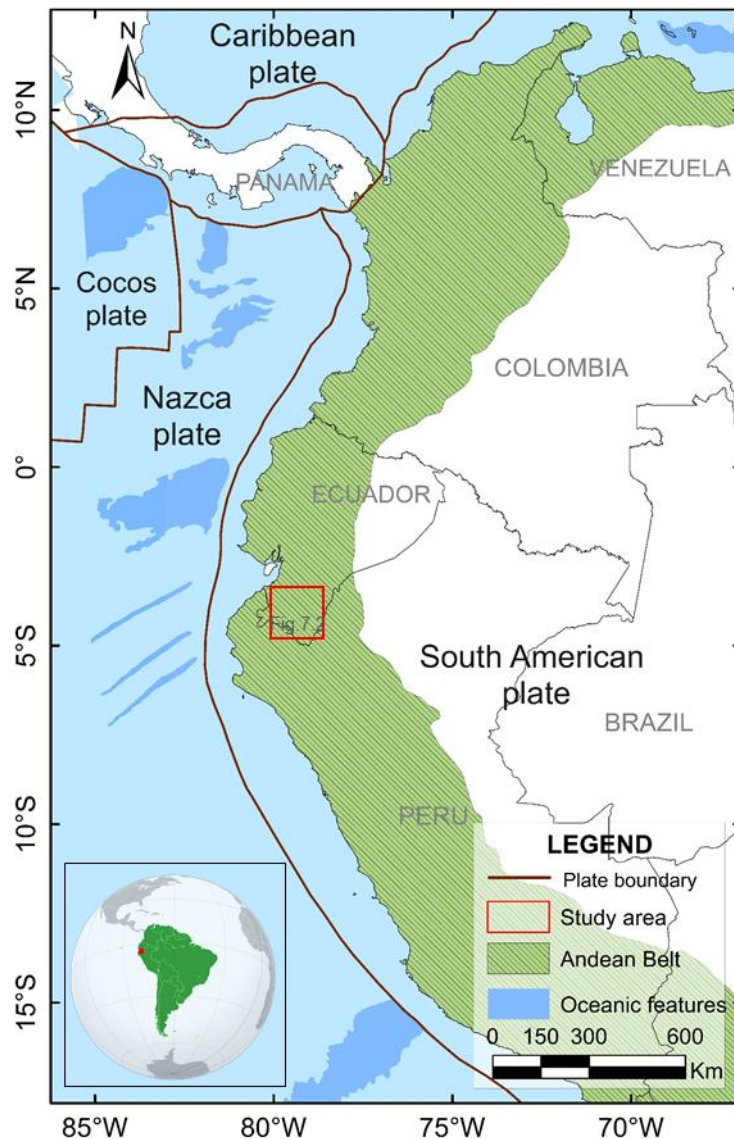


Figure 7.1 Simplified sketch of the South American Plate highlighting major plate boundaries and the location of the study area

Pecube is a 3D thermo-kinematic modeling code that enables the simultaneous analysis of multiple thermochronological datasets across different tectonic blocks. This capability is particularly important in geologically complex regions, where exhumation, fault activity, and topographic evolution are spatially heterogeneous and dynamically linked. Pecube integrates the effects of

topography, differential faulting, and rock advection, allowing it to predict thermochronometer ages and cooling paths based on user-defined tectonic scenarios, thus overcoming the limitations of traditional closure-temperature-based models (Braun, 2003; Braun et al., 2006; Braun et al., 2012). In parallel, 4DTherm code is particularly useful for distinguishing between cooling related to magmatic emplacement and subsequent exhumation, especially in cases where U–Pb and AFT ages are similar.

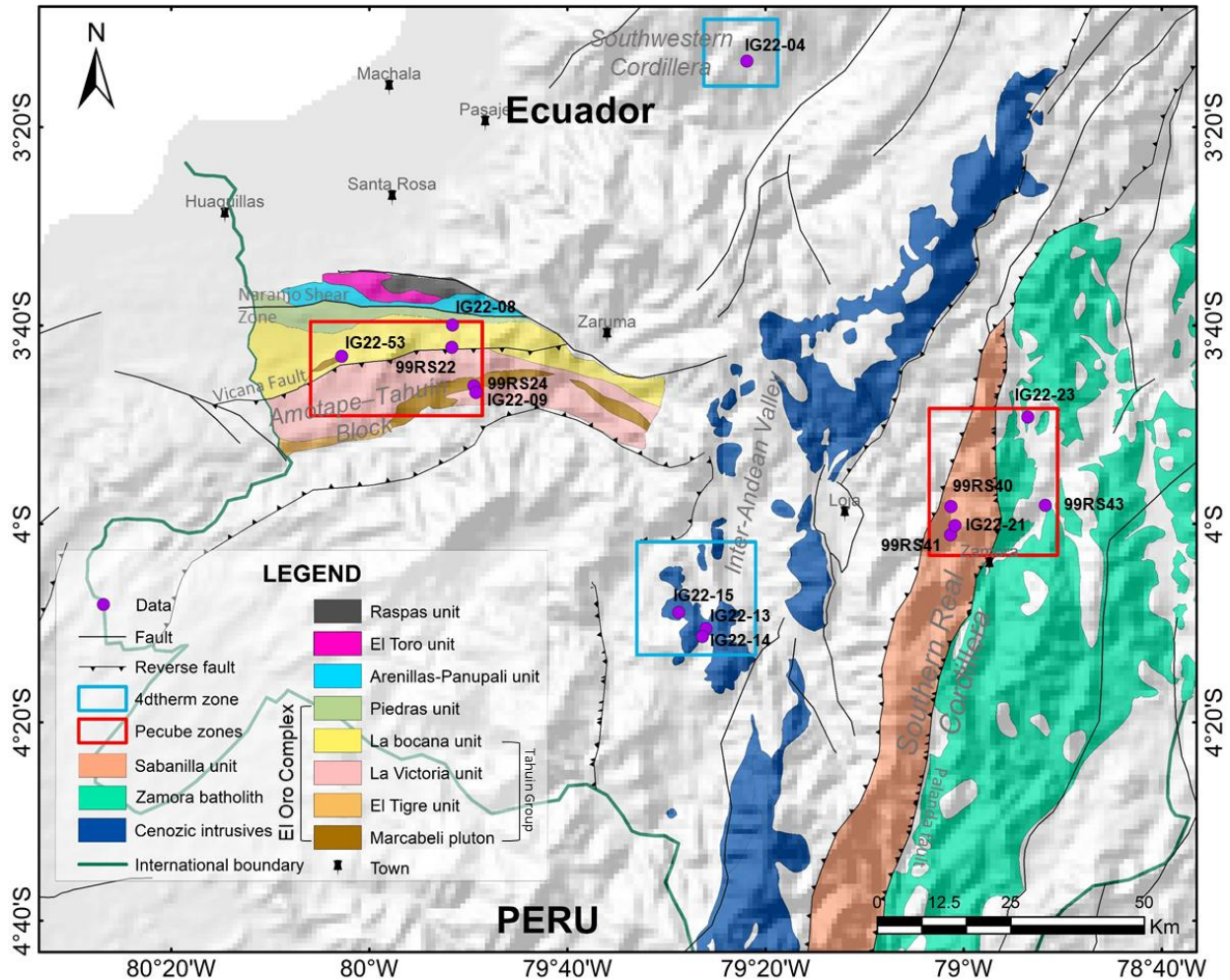


Figure 7.2 Geological generalized map of the study area showing the regions used for Pecube and 4DTherm modeling, and the locations of samples employed in the modeling.

In this study, we use Pecube to simulate the thermal evolution of southern Ecuador assuming steady-state topography. Our primary objective is to reconstruct the most plausible exhumation histories for two structurally complex regions (Fig. 7.2): (i) the Southern part of the Real Cordillera (SRC), which includes the Zamora Batholith and the Sabanilla Unit and is transected by the Palanda Fault; and (ii) the Amotape-Tahuin Block (ATB), which comprises the La Bocana and La Victoria units, juxtaposed along a major thrust fault. Additionally, we present 4DTherm modeling for samples that show

overlapping U–Pb and AFT ages (Chapter 6), to differentiate between magmatic, post-magmatic and erosional cooling. This integrated modeling approach provides new quantitative constraints on the timing and rates of exhumation in this key segment of the Northern Andes, enhancing our understanding of its tectono-thermal evolution.

7.2 Geological background

The Northern Andes, located in South America, are separated from the Central Andes by the Huancabamba Deflection at approximately 6°S (Mourier, 1988). Based on geological characteristics and major structural features, the Northern Andes are subdivided into three main segments: the Venezuelan Andes in the north, the Colombian Andes in the central portion, and the Ecuadorian-north Peruvian Andes to the south (Ramos & Aleman, 2000).

The southern Ecuadorian Andes are also divided into distinct geological domains, predominantly oriented north–south. In the west, the forearc zone encompasses low-relief basins and the ATB, which trends east–west and shifts to a north–south orientation in northern Peru. It comprises three juxtaposed geological complexes (Riel et al., 2014), including the El Oro Complex, which contains the Las Piedras Unit and the Tahuin Group (Fig. 7.1). The Tahuin Group consists of Devonian sediments of the El Tigre Unit (Martinez, 1970), Triassic granitoids of the Marcabeli and El Prado plutons (201–228 Ma; Litherland, 1994; Noble et al., 1997), metasedimentary rocks of the La Victoria Unit (0.2–1 Ga; IG22-09; Feininger & Silberman, 1982) and the Triassic migmatitic complex of the La Bocana Unit (Litherland, 1994), which hosts the La Florida Pluton (235 Ma; Sandoval-Espinel et al., 2025). The contact between the La Bocana and La Victoria units likely reflects a structure formed during the Late Triassic extensional event in the El Oro Complex (Riel et al., 2014). Although it is unclear whether a discrete normal fault originally developed along this specific contact, it is now expressed as a thrust fault, probably indicating subsequent tectonic inversion (Gómez Tapias et al., 2019). This structure is referred to here as the Vicana Fault.

Further east, the Celica–Lancones Basin is composed of igneous rocks related to a Late Cretaceous magmatic arc, along with volcanoclastic and siliciclastic sequences (Jaillard et al., 1999; Winter, 2008). Adjacent to this, the arc domain is represented by the Inter-Andean Valley, which is filled with volcanic products derived from Cenozoic to Recent magmatic activity (Jaillard, 2022).

Eastward, a high-relief morphotectonic structure is represented by the SRC, composed of Paleozoic to Lower Cretaceous metamorphic rocks (Litherland, 1994), including the Triassic Sabanilla Unit and the Jurassic calc-alkaline Zamora Batholith, juxtaposed along the Palanda Fault (Litherland, 1994; Pratt et al., 2005).

Further inland, the backarc zone encompasses the lowlands, which form the Amazonian piedmont of the Andes. This zone includes the Subandean fold-and-thrust belt and the adjacent Amazonian Basin.

7.3 Methodology

7.3.1 3D Thermokinematic-modelling (Pecube)

To reconstruct the thermal and exhumation history of the target areas, we employed the three-dimensional thermo-kinematic modeling code Pecube (Braun, 2003, 2012). This code uses a mixed Eulerian–Lagrangian approach and is designed to solve the three-dimensional heat transfer equation in a crustal block, incorporating all three advection terms (Eq. 7.1).

$$\frac{\partial T}{\partial t} + u \frac{\partial T}{\partial x} + v \frac{\partial T}{\partial y} + w \frac{\partial T}{\partial z} = \frac{\partial}{\partial x} \kappa \frac{\partial T}{\partial x} + \frac{\partial}{\partial y} \kappa \frac{\partial T}{\partial y} + \frac{\partial}{\partial z} \kappa \frac{\partial T}{\partial z} + A \quad \text{Equation 7.1}$$

where u, v, w are the component of a general velocity field representing the relative motion of rock with respect to a fixed base. A is Heat production. κ is thermal diffusivity.

Pecube is coupled with a Neighborhood Algorithm (see Sambridge, 1999) to explore a range of tectono-geomorphic scenarios, including exhumation and/or relief evolution, and thermo-kinematic parameters within predefined a priori bounds. The algorithm iteratively samples the model space to identify parameter combinations that minimize the misfit (Eq. 7.2) between observed and predicted thermochronological data.

$$\Phi = \sum_{i=1}^N \left(\frac{\alpha_{imod} - \alpha_{idat}}{\sigma_i} \right)^2 \quad \text{Equation 7.2}$$

We constrained the exhumation history of two regions in southern Ecuador: (i) the SRC, comprising the Sabanilla Unit and the Zamora Batholith; and (ii) the La Bocana and La Victoria units of the ATB. These regions were selected based on the availability of thermochronological data (Table 7.1), discrepancies noted in previous estimates of cooling rates, and the advantageous orientation of sample transects, which are approximately perpendicular to the main topographic relief. The exhumation history was modeled as a series of discrete phases, each defined by a constant exhumation rate between specified start and end times. To optimize computational efficiency, the topographic surface was simplified by resampling the Shuttle Radar Topography Mission (SRTM) digital elevation model to a resolution of 1 km—sufficient for accurate modeling results, as demonstrated by Valla et al. (2011).

Table 7.1 Summarized thermochronological data used in this study

Sample name	Long.	Latit.	Altitude	Thermochronological method		Source	
				AHe age	1 σ -Error		
99RS43	-78.863	-3.969	875	7.0	0.1		Spikings et al. (2010)
99RS40	-79.022	-3.971	1974	43.6	0.6		Spikings et al. (2010)
99RS41	-79.022	-4.018	1500	38.3	1.0		Spikings et al. (2010)
				AFT age	1 σ -Error	MTL +/- 1 σ -Error	
99RS24	-79.824	-3.768	685	45.6	3	12.7 $\pm$ 0.2	Spikings et al. (2005)
99RS22	-79.861	-3.703	0	38.8	2.7	13.6 $\pm$ 0.2	Spikings et al. (2005)
IG22-08	-79.860	-3.666	226.14	40.3	2.5	13.2 $\pm$ 1.0	This study
IG22-09	-79.825	-3.768	749.56	25.5	1.8	13.7 $\pm$ 1.4	This study
IG22-53	-80.046	-3.718	321.3	35.5	2.4	13.3 $\pm$ 1.4	This study
99RS43	-78.863	-3.969	875	34.0	3.3	12.6 $\pm$ 0.5	Spikings et al. (2010)
99RS40	-79.022	-3.971	1974	45.0	2.0	13.2 $\pm$ 0.2	Spikings et al. (2010)
99RS41	-79.022	-4.018	1500	39.5	2.0	12.5 $\pm$ 0.2	Spikings et al. (2010)
IG22-21	-79.015	-4.003	1546	36.2	2.6	13.1 $\pm$ 1.6	This study
IG22-23	-78.892	-3.820	874	31.3	2.8	13.9 $\pm$ 0.7	This study
				ZFT age	1 σ -Error		
99RS24	-79.824	-3.768	685	137.0	17.0		Spikings et al. (2005)
99RS22	-79.861	-3.703	0	180.0	17.0		Spikings et al. (2005)
99RS43	-78.863	-3.969	875	101.0	5.3		Spikings et al. (2010)
99RS40	-79.022	-3.971	1974	57.9	3.0		Spikings et al. (2010)
99RS41	-79.022	-4.018	1500	61.2	3.3		Spikings et al. (2010)
				⁴⁰ Ar/ ³⁹ Ar age (Muscovite)	1 σ -Error		
09RC40	-79.852	-3.704	461	214.0	0.9		Cochrane et al. (2014)
				⁴⁰ Ar/ ³⁹ Ar age (biotite)	1 σ -Error		
99RS40	-79.022	-3.971	1974	73.9	0.3		Spikings et al. (2010)
				⁴⁰ Ar/ ³⁹ Ar age (Hornblende)	1 σ -Error		
99RS43	-78.863	-3.969	875	174.0	1.0		Spikings et al. (2010)

The model setup for the SRC zone comprised a rectangular domain measuring 29×37 km (see Fig. 7.2), with a vertical fault (90° dip) of 15 km depth positioned outside the modeled block, to model the exhumation of rocks. The boundary conditions were a basal temperature ranging from 300 to 900 $^\circ\text{C}$, a surface temperature of 24 $^\circ\text{C}$ and an atmospheric lapse rate of 6 $^\circ\text{C}/\text{km}$. The heat production rate was permitted to vary between 0 and 100 $^\circ\text{C}/\text{My}$, while the model thickness was fixed at 55 km (Araujo, 2016). The initial configuration is illustrated in Fig. 7.3A.

During our modeling processes, positive rates (+) are interpreted as periods of exhumation and consequent cooling. While, negative rates (–) signify more complex thermal scenarios where samples

experienced apparent reheating, thermal overprinting or isotherm relaxations. This last can be potentially driven by magmatic heating, lithospheric thinning, or burial episodes.

The inversions were run with one to four phases of exhumation ($Pe = 1$ to 4). The first pulse was constrained to be positive, representing a phase of forced exhumation. The subsequent three pulses were modeled with variable rates ranging from -2 to 2 km/Ma (either positive related to exhumation, or negative associated with thermal relaxation). The model is additive, meaning that exhumation rates from the second pulse onward build upon those of the preceding phases. In this framework, positive rates (+) are interpreted as periods of exhumation and consequent cooling. While, negative rates (–) signify more complex thermal scenarios where samples experienced apparent reheating. This typically results from the relaxation of isotherms towards steady-state following a thermal perturbation, potentially driven by magmatic heating, lithospheric thinning, or burial episodes.

For the ATB zone, the model comprises a rectangular area measuring 37×23 km (see Fig. 7.2), incorporating the Vicana Fault at the center of the block, extending down to a maximum depth of 15 km, this fault separates the La Bocana unit (Block 1) from the La Victoria unit (Block 2). The fault is represented with a 30° dip, consistent with a moderate-angle thrust geometry. Although the possibility of a pre-existing normal fault could imply a steeper dip, 30° was selected as it corresponds to the mechanically optimal angle for thrusting, following Andersonian fault theory (Anderson, 1905). The thermal boundary conditions were identical to those of the SRC zone: a basal temperature range of 300 – 900 °C, a surface temperature of 24 °C and a lapse rate of 6 °C/km. The heat production range remained at 0 – 100 °C/My while the model thickness was fixed at 45 km (Araujo, 2016). The initial configuration is illustrated in Fig. 7.3B. Inversions were run with one to four phases of exhumation in each block ($Pe = 1$ – 1 to 4 – 4), where the notation $Pe=x$ – y indicates the number of exhumation phases imposed in Block 1 (x) and Block 2 (y).

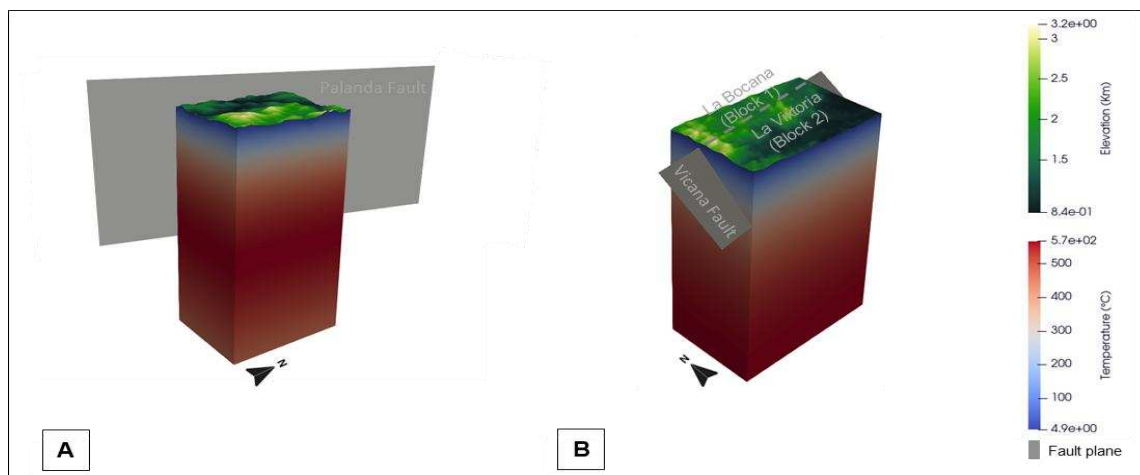


Figure 7.3 Initial set-up for Pecube modeling: (A) South Real Cordillera (SRC) and (B) Amotape–Tahuin Block (ATB).

Table 2.9 summarizes the fixed thermal and flexural parameters used in Pecube, while Table 2.10 lists the variable parameters.

To identify the optimal family of inverse models for the study areas, we calculated the Bayesian Information Criterion (BIC; Eq. 7.3; Braun et al., 2012; Neath & Cavanaugh, 2012). In addition, we quantified the search behavior (SB; Eq. 7.4), which evaluates the relative improvement of successive inversions. These two criteria allow us to balance goodness of fit and model complexity (via BIC) while ensuring that the inversions adequately explore parameter space (via SB), thereby reducing the risk of overfitting or bias toward local minima.

$$BIC = -2\ln(L) + k\ln(N) \quad \text{Equation 7.3}$$

where, k is the number of free parameters and N is the number of observations, respectively.

$$SB = \frac{\bar{m} - m_{best}}{\bar{m}} \quad \text{Equation 7.4}$$

Where, SR is the relative improvement of the inversion results, $\bar{m}$ is the mean misfit calculated across all tested models within a family and m_{best} is the lowest misfit.

7.3.2 2D Thermal and exhumation history of intrusions (4DTherm)

To quantify the exhumation histories of those samples which present nearly identical AFT and U–Pb zircon ages we employed the 4DTherm code (Fu et al., 2010). 4DTherm v.1.2 uses a 2D explicit finite difference solution which addresses conduction cooling, latent heat of crystallization/ fusion, convection within magma bodies, and hydrothermal circulation induced by magma intrusion as well as exhumation and erosion processes, enabling the reconstruction of thermal histories and distinguishing between magmatic, postmagmatic cooling and subsequent exhumation processes.

4DTherm solves the governing equations of heat and mass transfer in a magmatic body using an explicit finite difference scheme. The model accounts for conductive and convective cooling, as well as erosional exhumation. It conserves mass, momentum, and energy, under the following mathematical formulation:

$$\nabla(\rho_f U) = 0; \quad U = -\frac{k}{\mu_f}(\nabla\rho + \rho_f g); \quad \rho_b C_b \frac{\partial T}{\partial t} = \lambda_b \nabla^2 T - \nabla(\rho_f C_f U T) + Q \quad \text{Equation 7.5}$$

where symbols are defined in Table 2.11

The initial model setup is illustrated in Figure 7.6A. We iteratively ran models until achieving consistent results, with successful convergence in general matching mode. The geochronological data used in the models are presented in Table 7.2.

Table 7.2 Geochronological data used in the 4DTherm modelling

Sample Name	Latitude	Longitude	Elevation (m)	U-Pb age (Ma)	$\pm 1\sigma$ (Ma)	Source U-Pb age	AFT age (Ma)	$\pm 1\sigma$ (Ma)	Source AFT age
IG22-04	-3.2238	-79.3621	2246.37	29.7	2.0	Iglesias et al. (2025, submitted)	25.3	3.6	Chapter 6
IG22-13	-4.1758	-79.4336	2024.37	18.5	0.2	Iglesias et al. (2025, submitted)	19.9	3.6	Chapter 6
IG22-14	-4.1891	-79.4395	2133.67	21.7	0.4	Iglesias et al. (2025, submitted)	23.1	6.6	Chapter 6
IG22-15	-4.1486	-79.4793	1427.23	36.7	0.4	Iglesias et al. (2025, submitted)	38.0	17.0	Chapter 6

Additionally, Table 2.11 summarizes both the fixed boundary conditions and the variable parameters employed in the software.

7.4 Results

7.4.1 3D Thermal-kinematic modeling (Pecube)

For the SRC model, the lowest BIC value obtained was 370.1 (Table 7.3), corresponding to the $Pe = 4$ configuration. The SB value of 0.98 for this configuration indicates that the model family efficiently explored the parameter space while achieving a substantially better fit than the ensemble mean. The NA-plot dispersions (Fig. 7.4), together with the 1D and 2D Bayesian marginal distributions, show good convergence and mainly unimodal 1D distributions, supporting the selection of $Pe = 4$ as the most robust family of models. For this best-fit model, the mean basal temperature is 453.20 °C and the mean heat production is 14.02 m⁻³. The best-fit model within this family (lowest misfit) indicates four events: t_1 at 176 Ma with a rate (E_1) of -0.06 km/Ma; t_2 at 172 Ma with $E_2 = -1.6$ km/Ma; t_3 at 62.7 Ma with $E_3 = 0.2$ km/Ma; and t_4 at 9.1 Ma with $E_4 = 0.3$ km/Ma. The first rate (E_1) is too small to be considered a significant event and was therefore excluded from further interpretation.

For the ATB analysis, the lowest BIC (43.39) was obtained for the $Pe = 1-1$ configuration (one event per block). However, its SB value is negligible (0.05), indicating that the model explored only a very restricted portion of the parameter space and converged on a narrow set of solutions. In contrast, the $Pe = 4-4$ configuration (four events per block) yields the highest SB (0.99), reflecting a broader and more effective exploration of the parameter space. Examination of the 1D and 2D Bayesian marginal distributions across all model families (from $Pe = 1-1$ to $Pe = 4-4$) indicates that the $Pe = 4-4$ configuration achieves the most robust convergence, characterized by predominantly unimodal 1D distributions (Fig. 7.5).

Table 7.3 Summary of inversions performed for southern Ecuador. The preferred model is shown highlighted in gray color

Phase of exhumation (Pe)	Dimension (N)	$-2 \times \ln(L)$	Number of observations (k)	Number of models	BIC	SB
SRC						
1	4	2058.7	18	10100	2070.3	0.01
2	6	1035.5	18	22100	1052.8	0.11
3	8	582.5	18	35100	605.7	0.88
4	10	341.1	18	45100	370.1	0.98
ATB						
1-1	6	28.00	13	10100	43.39	0.05
1-2	8	27.77	13	25100	48.29	0.02
1-3	10	27.64	13	40100	53.29	0.30
1-4	10	63.52	13	60100	89.17	0.47
2-1	8	28.11	13	25100	48.63	0.19
2-2	10	26.87	13	45100	52.52	0.19
2-3	12	27.02	13	60100	57.80	0.69
2-4	14	27.07	13	75100	62.98	0.79
3-1	10	27.81	13	45100	53.46	0.13
3-2	12	29.45	13	60100	60.22	0.52
3-3	14	29.45	13	75100	65.36	0.88
3-4	16	28.34	13	90100	69.37	0.90
4-1	12	29.07	13	60100	59.85	0.34
4-2	14	28.57	13	70100	64.48	0.74
4-3	16	28.91	13	90100	69.95	0.76
4-4	18	29.06	13	120100	75.23	0.99

Based on these criteria, the best-fit model was selected from the Pe = 4–4 family. For clarity, we use the following notation: t_{Bn} and t_{Vn} denote the n-th exhumation or thermal phase in La Bocana and La Victoria Units, respectively, with the corresponding rates expressed as E_{Bn} and E_{Vn} . Positive rates indicate exhumation-driven cooling, whereas negative rates reflect thermal relaxation (isothermal rebound).

For this best-fit model, the mean basal temperature is 300.4 °C, and the mean heat production is 0.4 °C/My. For La Bocana, the model predicts: $t_{B1} = 244.4$ Ma with $E_{B1} = -0.6$ km/Ma; $t_{B2} = 95.4$ Ma with $E_{B2} = 1.7$ km/Ma; $t_{B3} = 67.1$ Ma with $E_{B3} = 1.2$ km/Ma; and $t_{B4} = 36.8$ Ma with $E_{B4} = 0.2$ km/Ma. For La Victoria, the predicted events are: $t_{V1} = 234.9$ Ma with $E_{V1} = -1.9$ km/Ma; $t_{V2} = 184.5$ Ma with $E_{V2} = 0.05$ km/Ma; $t_{V3} = 63.8$ Ma with $E_{V3} = 0.4$ km/Ma; and $t_{V4} = 23.5$ Ma with $E_{V4} = 0.3$ km/Ma. The negligible rate of E_{V2} does not represent a true geological event, and was therefore excluded from further interpretation.

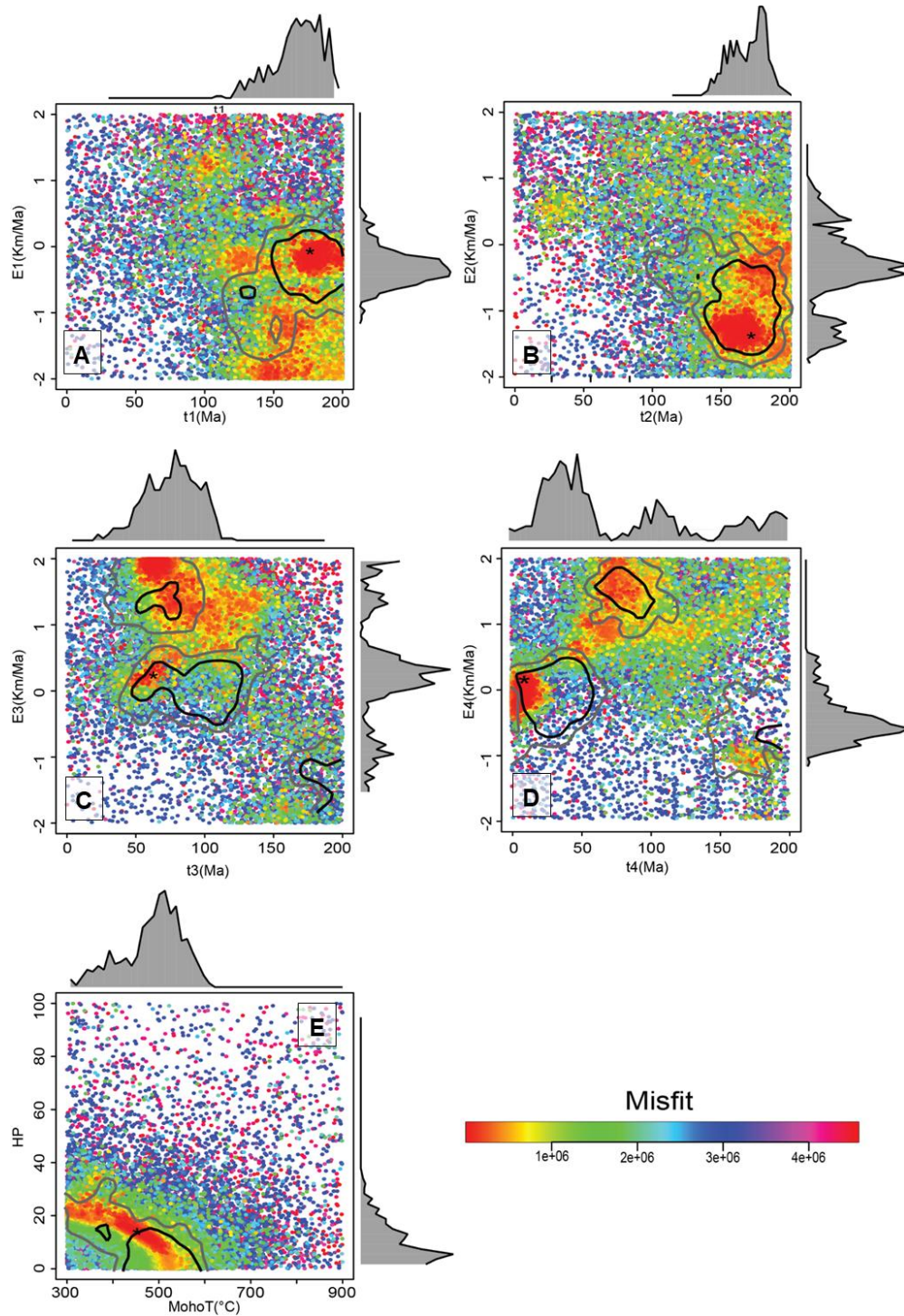
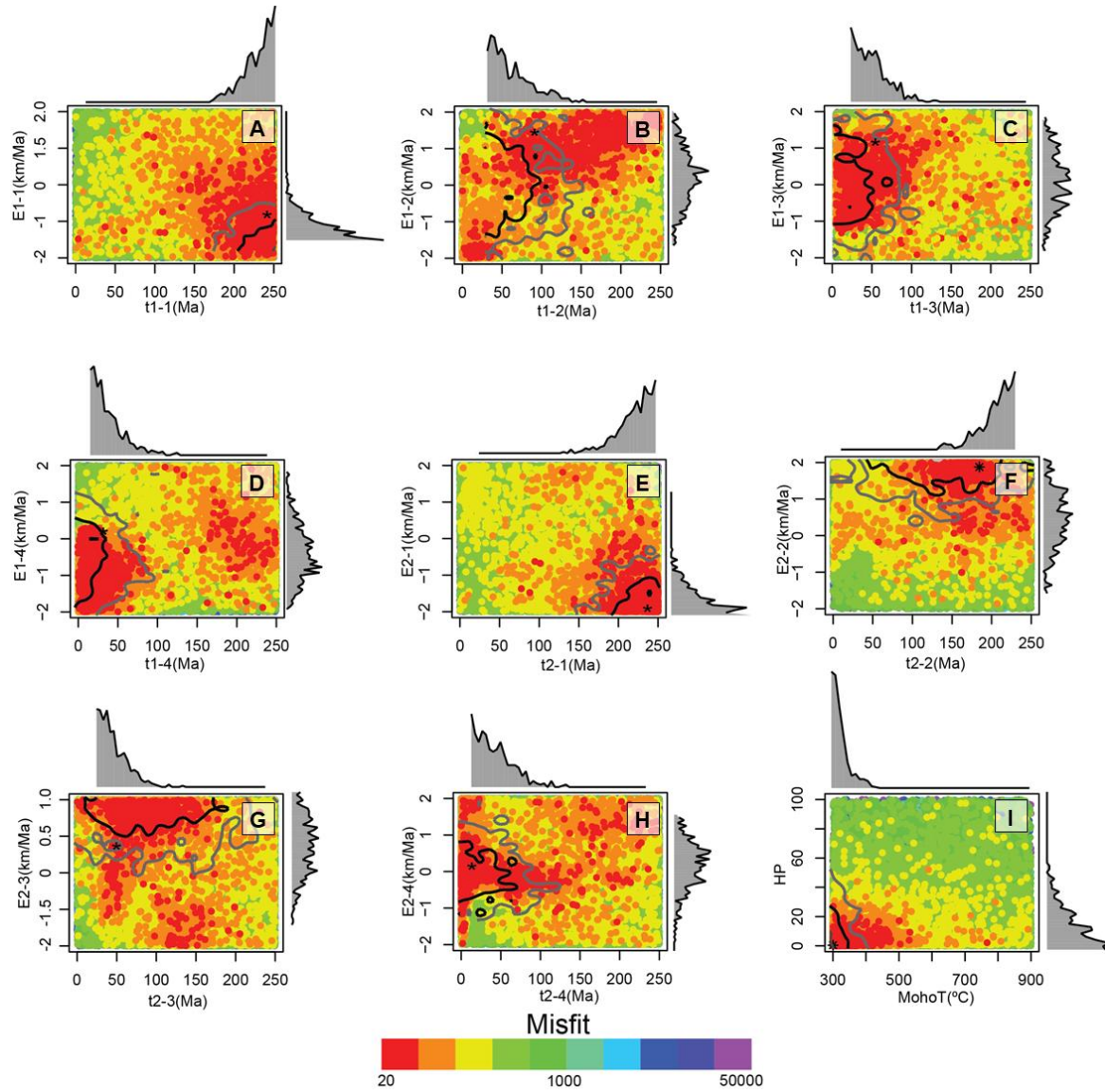


Figure 7.4 Results of the preferred thermo-kinematic model inversion assuming steady-state topography for the SRC. (A–E) Individual forward-model results (dots, colored according to misfit) and the corresponding 1D and 2D posterior probability-density functions (PDFs) of parameter values. (A) Onset time of exhumation versus initial exhumation rate. (B) Timing of the second exhumation pulse versus its exhumation rate. (C) Timing of the third exhumation pulse versus its exhumation rate. (D) Timing of the fourth exhumation pulse versus its exhumation rate. (E) Basal temperature versus radiogenic heat production (HP). The black asterisk marks the best-fit model (lowest misfit). One-dimensional marginal PDFs of parameter values from the NA-Bayes analysis are shown along the top and right axes of each plot, while the 2D marginal PDFs are shown as black (1σ) and grey (2σ) contours.



7.4.2 2D Thermal and exhumation history of intrusions (4DTherm)

The modeled cooling paths for samples IG22-04, IG22-13, IG22-14, and IG22-15 successfully reproduce all thermochronological constraints (Fig. 7.6).

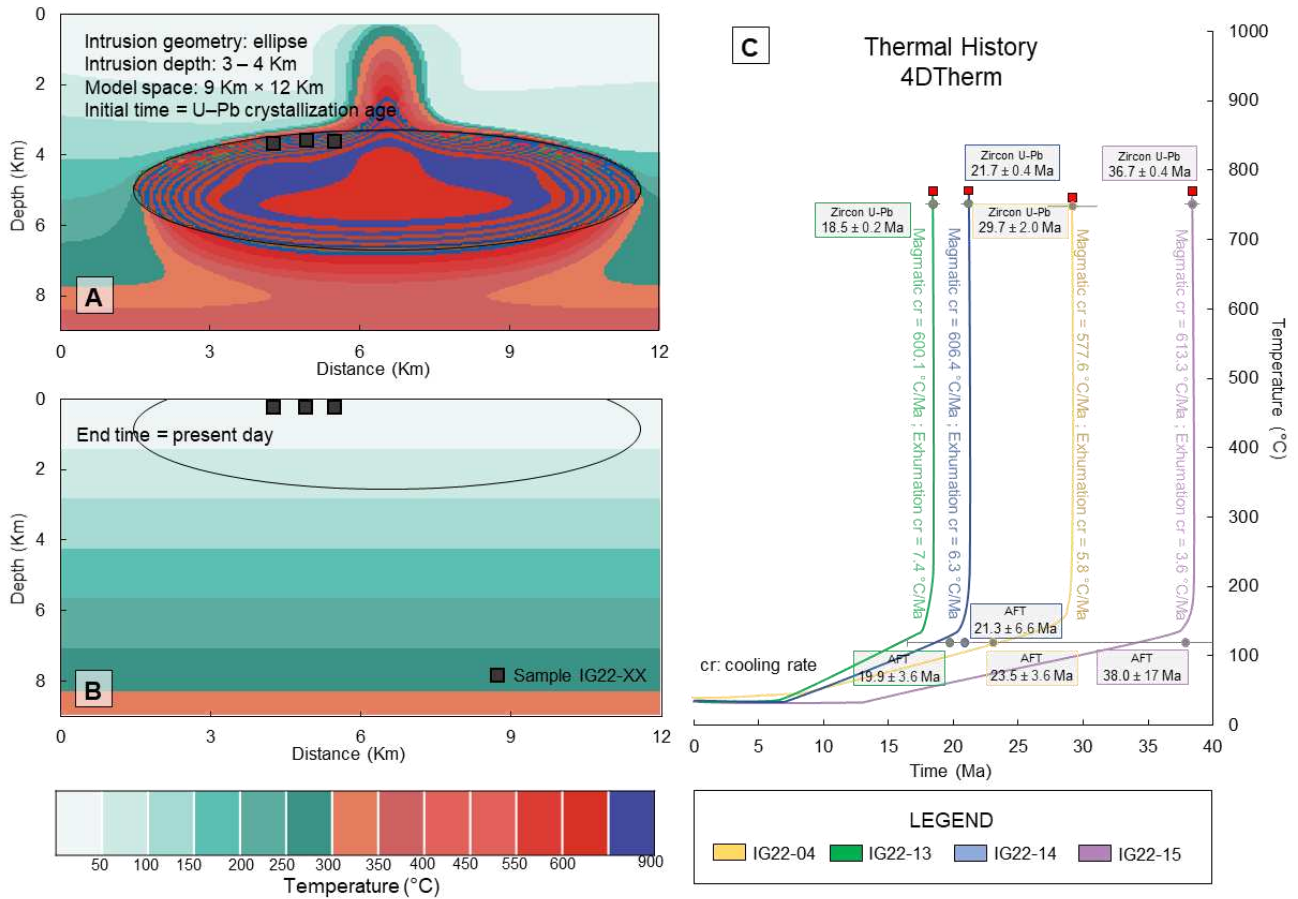


Figure 7.6 (A) General setup for the 4D-Thermo modeling at the initial time. (B) Visualization of the final stage of the 4D-Thermo modeling. (C) Time–temperature plot illustrating the cooling paths of the studied samples, derived from the 4D-Thermo modeling results.

Sample IG22-04, emplaced at ~3400 m at 29.7 Ma and cooled to 166.3 °C by 28.7 Ma. It records a rapid magmatic cooling rate of ~577.6 °C/Ma, followed by a slower exhumation-driven cooling rate of ~5.8 °C/Ma, corresponding to an average exhumation rate of ~0.12 km/Ma. For IG22-13, emplacement occurred at a depth of ~3000 m at 18.5 Ma, followed by cooling to 129.3 °C by 17.45 Ma, with estimated magmatic and exhumation-related cooling rates of 600 °C/Ma and 7.4 °C/Ma, respectively, and an average exhumation rate of 0.16 Km/Ma. Similarly, sample IG22-14 was

emplaced at approximately 3100 m at 21.7 Ma and cooled to 130.8 °C by 20.6 Ma, with corresponding cooling rates of 606.4 °C/Ma (magmatic) and 6.33 °C/Ma (exhumation), and an average exhumation rate of 0.14 Km/Ma. For IG22-15, emplacement occurred between ~3200 and 3500 m at 36.7 Ma, reaching 134.2 °C by 35.68 Ma. In this case, magmatic cooling was 613.24 °C/Ma, and exhumation cooling was estimated at 3.56 °C/Ma, with an exhumation rate of 0.07 Km/Ma.

7.5 Discussion

7.5.1 Reliability of Pecube modeling results

Thermokinematic modeling for the SRC region reveals a good correlation between predicted and observed thermochronological ages (Fig. 7.7A), with most predictions falling within the 2σ uncertainty range of the data. The model reliably captures mid- to high-temperature cooling histories, as reflected by the close agreement between predicted and measured $^{40}\text{Ar}/^{39}\text{Ar}$ ages of biotite and hornblende. In contrast, AFT ages display greater variability: while some samples (e.g., 99RS40, measured 45 ± 2.0 Ma; predicted 49.2 Ma) show close correspondence, others (e.g., IG22-23, measured 31.3 ± 2.8 Ma; predicted 47 Ma) indicate moderate divergence. ZFT ages exhibit the largest discrepancies, with notable cases such as 99RS43 (measured 101 ± 5.3 Ma; predicted 80.4 Ma) highlighting potential underestimation of early cooling events. These mismatches likely reflect oversimplified boundary conditions in the current model setup, specifically: (1) the assumption of steady-state topography, and (2) the role of the Palanda fault, which plausibly partitioned the region into domains with distinct uplift and erosion histories. For instance, on the footwall side, model–data agreement is generally closer, whereas samples from the hanging wall tend to show larger deviations.

The ATB model yields an overall stronger fit (Fig. 7.7B), reproducing both AFT and ZFT ages within analytical uncertainties, and indicating a robust capture of low- to intermediate-temperature cooling. However, $^{40}\text{Ar}/^{39}\text{Ar}$ muscovite ages are systematically underpredicted, possibly due to argon loss through diffusion or late-stage thermal overprinting. Alternatively, the mismatch may reflect limitations in the thermal–kinematic setup, parameterization, or spatial resolution of the model.

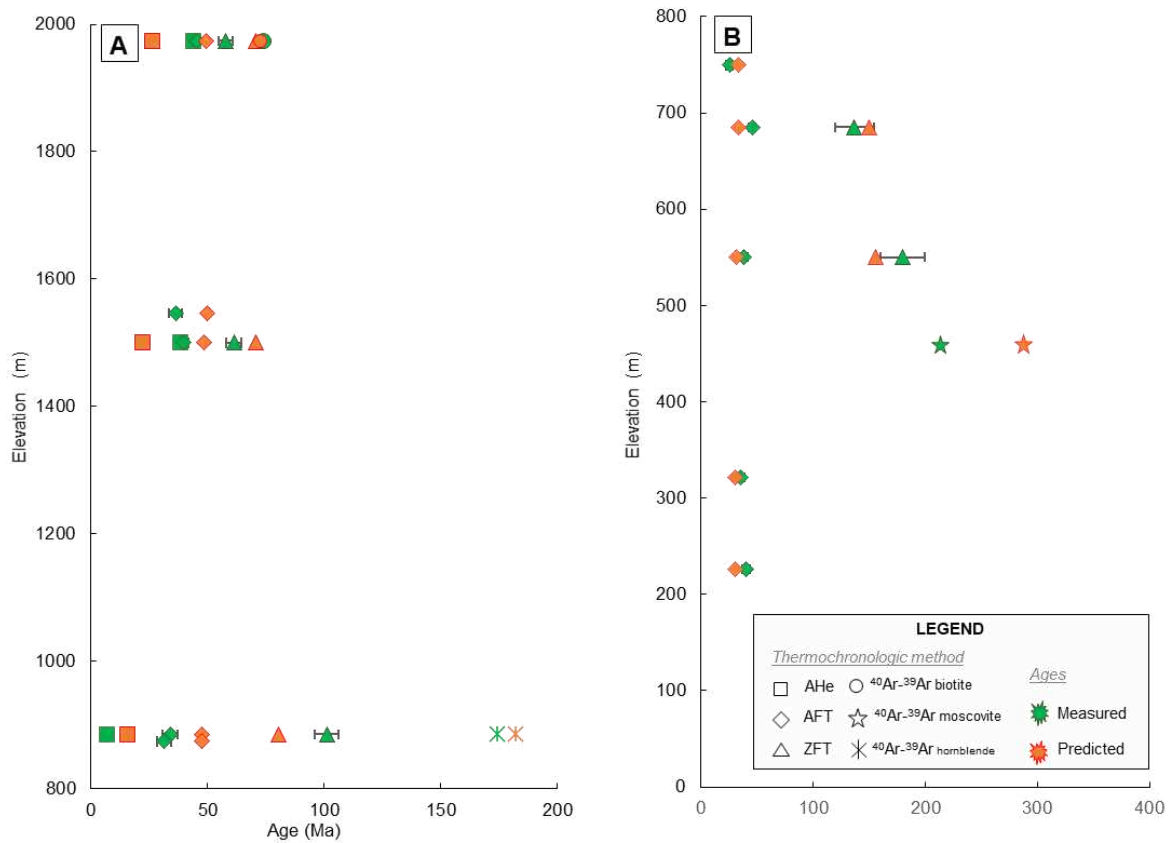


Figure 7.7 Age–elevation plots comparing measured and predicted data for (A) the SRC and (B) the ATB.

The reliability of both models is further supported by the unimodal nature of the 1D Bayesian marginal distributions for most inverted parameters (Figs. 7.4 and 7.5), including exhumation timing, rates, basal temperature, and heat production. This parameter convergence suggests minimal trade-offs in the inversion process, reinforcing the robustness of the reconstructed thermal histories for both the SRC and ATB regions.

7.5.2 2D Thermal and exhumation history of Cenozoic intrusions (4DTherm)

The modeled thermal histories of samples IG22-04, IG22-13, IG22-14, and IG22-15 suggest emplacement depths of about 3000–3500 meters. This is consistent with relatively shallow pluton intrusion in the upper crust (Burov et al., 2003). These samples exhibit a similar thermal regime, which is characterized by rapid magmatic cooling (613–577 °C/Ma) and more protracted cooling associated with exhumation (7.4–3.6 °C/Ma). Comparable patterns of shallow emplacement and rapid post-emplacement cooling have also been reported for several Andean plutonic complexes (e.g., Assis et al., 2022; Campos et al., 2009; Makshev et al., 2009).

The shallow emplacement depths and rapid cooling recorded in the Late Eocene–Early Miocene samples suggest a tectonic regime favoring upper-crustal magma storage. Such conditions are typical

of extensional settings, where crustal thinning reduces lithostatic pressure and promotes magma ascent. However, compressional regimes can also facilitate shallow intrusion through crustal thickening and fault-controlled pathways, while sediment subduction contributes to the generation of buoyant, silica-rich magmas capable of rapid ascent and crystallization at shallow depths.

During the Late Eocene (45–30 Ma), zircon isotopic signatures reported by Witt et al. (2023) suggest a broad extensional magmatic arc, consistent with regional crustal thinning. This interpretation is further supported by (Tamay-Granda, 2018), who documented syn-magmatic extensional basin development in southern Ecuador through integrated geological and geophysical observations. Such an extensional regime would have promoted efficient magma ascent through thinned lithosphere and favored the formation of shallow-crustal plutons. In contrast, Early Miocene magmatism appears linked to a compressional regime dominated by sediment subduction (Witt & Bourgois, 2010), which would have generated buoyant, silica-rich melts through slab-derived flux melting. The resulted hydrous, evolved magmas could rise rapidly and crystallize at shallow crustal levels.

7.5.3 Exhumation history and regional tectonic evolution

Triassic – Jurassic

The Triassic–Jurassic tectonic evolution of southern Ecuador is widely attributed to lithospheric thinning and extension, (Chapter 4; Riel et al., 2018; Spikings et al., 2015; Villares et al., 2020). Our 3D thermo-kinematic models (t_2 , t_{B1} , t_{V1}) reveal intervals of negative exhumation rates, which we interpret as signatures of isothermal rebound and lithospheric thermal relaxation. The temporal correspondence between these modeled signals and independent evidence for Jurassic extension suggests that crustal thinning progressively accompanied this phase.

Albian – Coniacian (101 – 88 Ma)

During the Albian (~101 Ma), active seafloor spreading in the South Atlantic and the westward drift of the South American Plate (Eagles, 2007) drove a regional tectonic reorganization. This transition marked the shift from a predominantly extensional regime to a compressional Andean-type subduction system, leading to the establishment of a new continental arc along the margin (101–88 Ma; Chapter 4; Ramos & Aleman, 2000; Winter, 2008). In this framework, the exhumation event at ca. 95 Ma recorded in the La Bocana unit likely represents the initial response to the subduction-driven geodynamic transition.

Maastrichtian – Middle Paleocene (~70–60 Ma)

The Paleocene represents a phase of rapid uplift in the Ecuadorian Andes (Spikings et al., 2005, 2010; Vallejo et al., 2009, 2019), coinciding with a well-documented thermal anomaly dated between ~70 and 60 Ma (Litherland, 1994). Geological evidence for this event includes the arrival of detrital quartz (~70 Ma) on the eastern flank of the present-day Western Cordillera and a pronounced sedimentary hiatus along the continental margin (Jaillard, 2025). Our Pecube thermal–kinematic models capture this signal, identifying exhumation events at 67.1 Ma (t_{B3}), 63.8 (t_{V3}) and 62.7 Ma (t_3).

Two tectonic mechanisms likely contributed to this event: (1) the Late Cretaceous–Paleocene accretion of oceanic terranes along the northwestern margin of South America (~75–65 Ma; Jaillard et al., 2009; Spikings et al., 2010; Vallejo et al., 2019; Witt et al., 2019), and (2) a reorganization of plate convergence dynamics, marked by a shift in Farallón plate motion from predominantly N or NW to NNE or NE at ~67 Ma (Seton et al., 2012) or ~60 Ma (Jaillard, 2025), which altered the regional stress field (Pardo-Casas & Molnar, 1987). Regardless of the primary driver, the Paleocene event stands out as a regionally pervasive geodynamic signal recorded across all studied domains.

Middle Eocene – Early Oligocene (~40–35 Ma)

The Late Eocene is widely recognized as a significant compressional phase in Ecuador (Jaillard et al., 1995; Naeser et al., 1991). This period was characterized by widespread erosion and a shift from marine to terrestrial deposition (Espurt et al., 2018; Witt et al., 2019). This period coincided with the deformation and emergence of the forearc basins (Jaillard, 1997), as well as a moderate increase in convergence rates (Maloney et al., 2013). In the La Bocana unit of the ATB, we identify an exhumation event at ~36.8 Ma, with an average rate of ~0.2 km/Ma. This timing likely coincides with previous estimates (~43–39 Ma; Spikings et al., 2005) (~45–38 Ma; Chapter 6), but remains consistent with a cooling rate of ~4.5 °C/Ma (Spikings et al., 2005) when a 35 °C/km geothermal gradient is applied (Angulo-Romero et al., 2023). Furthermore, although thermal modeling (4DTherm) indicates some exhumation-driven cooling in the Inter- Andean Valley during this interval (sample IG22-15; ~3.6 °C/Ma), the associated exhumation rate is extremely low (~0.07 km/Ma). This suggests that this may not represent a significant geologic exhumation event, but rather a period of tectonic quiescence in this crustal domain.

It is noteworthy that neither the 45–35 Ma exhumation in the SRC reported by Spikings et al. (2005) nor the 34–31 Ma cooling documented in Chapter 6 were resolved in our model for this crustal domain, likely reflecting a combination of modeling choices and data limitations. In our SRC

inversions, the area was represented as a single block with the principal fault placed outside the domain due to limit data, and steady-state topography was enforced; these simplifications may bias inversions toward spatially averaged and temporally smoother cooling histories, which could obscure short, fault-localized pulses.

Late Oligocene – Early Miocene (~29–18 Ma)

Around 25 Ma, fragmentation of the Farallón Plate triggered a major reorganization of plate boundaries (Jaillard, 2025; Maloney et al., 2013) and a shift in the subduction vector from ESE to E (Gordon & Jurdy, 1986; Pilger 1984). This tectonic change likely drove the Late Oligocene exhumation event recorded in the La Victoria unit of the ATB (Block 2) consistent with the Late Oligocene–Early Miocene timing (~27–22 Ma) reported in Chapter 6 but inconsistent with the uniform cooling between ~43 and 39 Ma reported by Spikings et al. (2005) for units south of the Naranjo Shear Zone. The contrasting exhumation histories of the La Victoria (~23.5 Ma) and La Bocana (~36.8 Ma) units further suggest that the Vicana fault (Fig. 7.2) exerted a significant control on uplift, supporting the interpretation proposed in Chapter 6 and challenging the notion of synchronous cooling across the ATB.

This exhumation phase coincides with the proposed Late Oligocene–Early Miocene magmatic flare-up in Ecuador (Schütte et al., 2010). Plutonic bodies emplaced during this interval (samples IG22-13, and IG22-14) intruded at relatively shallow crustal levels (~3.0–3.5 km), as constrained by our thermal modeling. Geochemical data from the nearby Portachuela Batholith, reported by Iglesias et al. (Submitted), indicate that magmatism in southern Ecuador between ~31–16 Ma was derived predominantly from an enriched mantle source with minimal crustal assimilation, an enrichment attributed to sediment subduction. The incorporation of silica- and volatile-rich components into the mantle wedge would have generated more evolved, buoyant, and water-rich magmas, conditions favorable for the emplacement of plutons at shallow crustal levels. Magmatic addition likely contributed to crustal thickening, with isostatic compensation driving surface uplift. In this context, the Interandean Valley may have begun exhuming during this interval. Samples IG22-14 and IG22-13 record higher exhumation rates (~0.14–0.16 km/Ma) compared to sample IG22-15 (~0.07 km/Ma at 36 Ma), the latter interpreted as reflecting an earlier phase of relative tectonic quiescence. We therefore suggest that the onset of Valley exhumation was likely influenced, at least in part, by magmatic addition and associated crustal thickening.

Late Miocene (~10–9 Ma)

By the Late Miocene, convergence between the Nazca (formerly Farallón) and South American plates had become nearly orthogonal, enhancing interplate coupling and driving widespread upper-plate shortening (Jaillard, 2025). This tectonic configuration triggered the ~9 Ma uplift recorded in the SRC (this study), coinciding with a regional angular unconformity dated to ~10–9 Ma, the shift from marine–tidal to fully continental sedimentation (Hungerbühler et al., 2002), and magmatism that contributed to substantial crustal thickening, including the development of adakite-like signatures (Iglesias et al. (Submitted)). We interpret this event as a major phase of topographic rise during the ongoing uplift of the Andes. Notably, the ~9 Ma pulse is clearly evident in our 3D inversions, yet it was not identified in earlier 2D thermochronological models (Chapter 6; Spikings et al., 2005).

7.6 Conclusions

Thermokinematic modeling (Pecube) of the Real Cordillera and the Amotape-Tahuin Block reveals Triassic-Jurassic events that can be best explained by isotherm relaxation associated with an extensional regime. Later events reveal distinct exhumation phases, most notably a strong, regionally consistent pulse from the Maastrichtian to the Early Paleocene (~67–63 Ma) that likely represents a significant geodynamic episode in southern Ecuador. The imperfect fit of some datasets (simulated versus measured) suggests that long-term, steady-state exhumation did not prevail in the region. Complementary 4DTherm simulations of the Western Cordillera and the Inter-Andean Valley provide evidence of shallow intrusive emplacement. The latter also records the possible onset of exhumation during the Late Oligocene–Miocene. Combining Pecube with 4D thermal modeling from different crustal domains demonstrates that southern Ecuador experienced a protracted and heterogeneous tectonothermal history.

Declaration of Competing Interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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CHAPTER 8: CONCLUSIONS

The southernmost segment of the Northern Andes, represents the transition zone between the Northern and Central Andes. This region, spanning southern Ecuador and northern Peru, is marked by a major change in overall orientation of the mountain chain from NE to NW, and preserves a nearly continuous record of Paleozoic to Miocene units. It also hosts exotic terranes that profoundly reshaped the arc and uniquely records both arc migration and basin development, a combination rarely observed elsewhere in the Andes. Because it integrates such a long and complex geological history, this area is key to how some subduction settings evolve. Despite its significance, this area has remained relatively understudied due to historical limitations in infrastructure and safety, resulting in scarce geological and thermochronological data. In this thesis, I address this gap by integrating new field observations with geochronological, isotopic, and thermochronological analyses. Specifically, I collected in situ and detrital samples for U–Pb zircon dating, Hf and O isotope measurements, trace element geochemistry, whole-rock analyses, apatite fission-track dating, and numerical modeling of intrusive rocks across southern Ecuador and northernmost Peru. Together, these datasets provide a more comprehensive understanding of the tectonic, magmatic, and exhumation history of this critical segment of the Andes.

From the present work we can broadly conclude

- 1) Detrital zircons spanning the Neoproterozoic to Triassic (1000–200 Ma) from forearc basins in southern Ecuador and northern Peru reveal two main age populations: a Late Neoproterozoic–Ordovician group (640–460 Ma) and a Permian–Triassic group (280–200 Ma). The geochemical dataset indicates that most zircons are products of crustal reworking, as evidenced by elevated $\delta^{18}\text{O}$ values, low ϵHf values, and trace element patterns characteristic of zircons derived from S-type granitoids. Provenance analysis suggests that these detrital zircons were sourced predominantly from northeastern Gondwana rather than displaying typical Andean signatures (i.e., central–southern Andes).
- 2) The geodynamic evolution of the Mesozoic arc in southern Ecuador and northern Peru developed within a predominantly extensional setting and is marked by a progressive transition from crustal-dominated to more mantle-like sources. This transition is recorded by a systematic increase in ϵHf values from the Triassic to the Early Cretaceous, in some cases approaching mantle-like signatures, together with a decrease in $\delta^{18}\text{O}$ values that converge toward the mantle array. By ~110 Ma, the arc migrated westward, a shift coeval with the opening of the Celica–Lancones Basin. This westward migration may have promoted crustal thinning, enabling the formation of an oceanic-like basement that subsequently hosted the Late Cretaceous arc. In addition, our new U–Pb zircon ages from northern Peru yielded Jurassic ages, challenging both the long-assumed absence of the Jurassic arc in this region and the proposed younging trend of Jurassic magmatism in the northern Andes.
- 3) The Cenozoic magmatic history of northern Peru is characterized by multiple overlapping granitoid intrusions, predominantly granodiorites and granites, emplaced between ~60 and 19 Ma. Notable examples include the Paltashaco pluton (Paleocene to Oligocene, 58.9–29.1 Ma) and the

Pamparumbe pluton (Eocene to Miocene, 43.6–19.9 Ma). Our new geochronological data redefine the timing of these intrusions, which were previously thought to be Late Cretaceous to Paleocene in age.

- 4) Apatite fission-track dating from southern Ecuador documents three cooling intervals: (i) middle Eocene (~45–38 Ma), (ii) late Eocene–early Oligocene (~40–30 Ma), and (iii) Oligocene (~33–25 Ma). These results demonstrate that exhumation in southern Ecuador began as early as the Middle Eocene, significantly predating the previously proposed onset at ~10–9 Ma. Two of the identified cooling phases (~45–38 Ma and ~33–25 Ma) broadly coincide with major reorganizations in Nazca–South America plate convergence, whereas the ~40–30 Ma event corresponds to a modest increase in convergence rate. Spatial variability in the timing and magnitude of cooling across distinct crustal blocks highlights the asynchronous nature of exhumation in this region. In addition, the overlap between U–Pb crystallization ages and AFT cooling ages in certain domains suggests strong post-magmatic cooling, reflecting rapid thermal relaxation following intrusion rather than exhumation-driven cooling. This interpretation is reinforced by 2D thermal and exhumation histories of Cenozoic intrusions, which indicate shallow emplacement depths and rapid subsequent cooling. Finally, 3D thermokinematic modeling (Pecube) of the Amotape–Tahuín block reveals that exhumation within the La Bocana and La Victoria units occurred at different times, suggesting spatially variable tectonic exhumation likely linked to fault activity, consistent with the Hefty modeling results.

In the future, several avenues of research remain to be explored. For example, incorporating lower-temperature apatite (U–Th)/He data, which are more sensitive to near-surface processes, would allow for a better understanding of topographic changes across the different tectonic blocks of southern Ecuador. In addition, apatite fission-track ages already reported for northern Peru should be interpreted in detail to provide insights into the exhumation history of that region. Another important step is to constrain variations in crustal thickness throughout the Mesozoic–Cenozoic, intervals characterized by significant changes in stress regimes and magma reservoirs.