

Tectono-stratigraphic and sedimentological analysis of the Early to Middle Devonian Floresta Formation: Insights from the Floresta Massif, Northern Andes, Colombia

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ABSTRACT. A comprehensive tectono-stratigraphic and sedimentological investigation of Early to Middle Devonian rocks was conducted in the southern Floresta Massif and adjacent regions in the Northern Andes of Colombia. A substantially reduced thickness of the Floresta Formation compared to prior studies is suggested here, attributable to pronounced stratal deformation and the prevalence of recumbent folds throughout the area. The deformation in the Floresta Formation manifests as atypical recumbent folds, diverging from the structural behavior observed in the underlying and overlying strata of the El Tibet and Cucho formations respectively, which exhibit minimal deformation. Our findings also reveal that the Floresta Formation accumulated under shallow-water platform conditions, subject to eustatic sea-level fluctuations. This resulted in distinct episodes of carbonate and siliciclastic deposition, with terrigenous sediments sourced from continental origins, potentially encompassing a combination of cratonic areas and uplifted blocks. The identification of a plausible stage of carbonate silicification signifies a post-diagenetic transformation. The sedimentary rocks of the Floresta Formation reached the upper epizone conditions, in proximity to the transition between the epizone and the upper anchizone, which suggests a maximum depth and temperature of ~5-7 km and ~300 °C, respectively. This contribution provides new insights into the geological history of the region, emphasizing the importance of scrutinizing Early to Middle Devonian rocks within the broader geological context of the Northern Andes.

Keywords: Floresta Massif, Diagenesis, Stratigraphy, Paleofauna, Deformation, Recumbent folds.

RESUMEN. Análisis tectono-estratigráfico y sedimentológico de la Formación Floresta (Devónico Temprano-Medio): Evidencias del Macizo de Floresta, Andes del Norte, Colombia. Este trabajo presenta una exhaustiva investigación tectono-estratigráfica y sedimentológica de las rocas del Devónico temprano a Medio en el sur del Macizo de Floresta y regiones adyacentes, Andes del Norte, Colombia. Se sugiere un espesor sustancialmente reducido para la Formación Floresta en comparación con estudios previos, atribuible a una deformación pronunciada de los estratos y a la prevalencia de pliegues recumbentes en toda el área. La deformación en la Formación Floresta se manifiesta mediante pliegues recumbentes atípicos, a diferencia del comportamiento estructural observado en los estratos subyacentes y sobreyacentes de las formaciones El Tibet y Cucho, respectivamente, que muestran una deformación mínima. Se revela también que la Formación Floresta se acumuló en condiciones de plataforma de aguas someras, sujeta a fluctuaciones eustáticas del nivel del mar. Esto resultó en variados episodios de depositación carbonática y siliciclástica, con aportes de sedimentos terrígenos provenientes de fuentes continentales, los que posiblemente abarcaron una combinación de áreas cratónicas y bloques levantados. La identificación de una posible etapa de silicificación carbonática indica una transformación

postdiagenética. Las rocas sedimentarias de la Formación Floresta alcanzaron las condiciones de la epizona superior, en proximidad a la transición entre la epizona y la anquizona superior, lo cual sugiere una profundidad y temperatura máximas de ~5-7 km y ~300 °C, respectivamente. Esta investigación aporta nueva información acerca de la compleja historia geológica de la región, enfatizando la importancia de examinar las rocas del Devónico Temprano a Medio en el contexto geológico más amplio de los Andes del Norte.

Palabras clave: Macizo de Floresta, Diagénesis, Estratigrafía, Paleofauna, Deformación, Pliegues recumbentes.

1. Introduction

In the northern Andes of Venezuela and Colombia, exposures of Devonian rocks, mostly of Emsian to Givetian ages (early to middle Devonian), usually occur in geographically limited areas widely separated from each other (Fig. 1). Other isolated outcrops are found along the Central Andes in southern Peru, Bolivia, Chile, and Argentina, and intracratonic basins in Brazil, Paraguay, Uruguay and Argentina (Barrett and Isaacson, 1988). In northwestern South America, Devonian rocks outcrop around some of

the ancient Quetame, Floresta and Santander massifs of the Colombian Andes, and in the Perijá Range in the northern region close to the boundary with the Caribbean tectonic domains (Morzadec *et al.*, 2015). The Floresta Massif has attracted attention in previous works of geological (*e.g.*, Ulloa and Rodríguez, 1982; Sotelo, 1997¹; Jiménez, 2000²; Ulloa *et al.*, 2003), paleontological (*e.g.*, Olsson and Caster, 1937; Caster, 1939; Royo y Gómez, 1942), sedimentological (*e.g.*, Botero, 1950), mineral

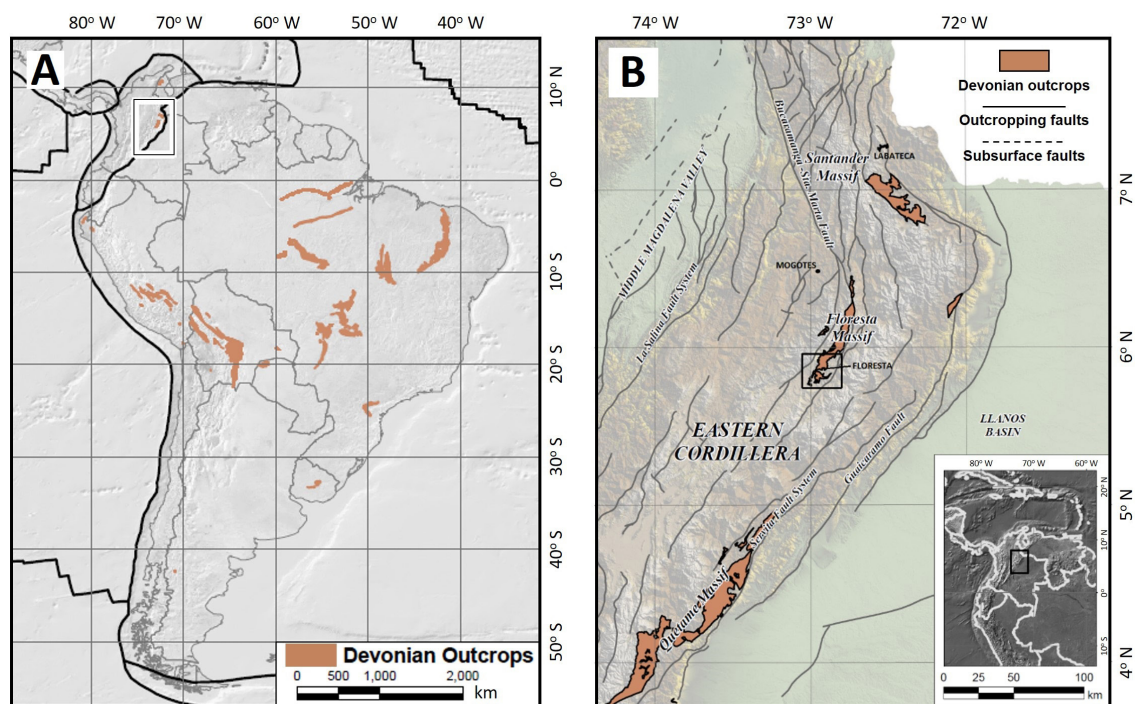


FIG. 1. **A.** Map of South America, showing Devonian outcrops and main tectonic plate boundaries. **B.** Zoom-in excerpt highlighting the Eastern Cordillera of Colombia (black square inset in A), showing the Quetame, Floresta and Santander massifs, and the main tectonic structures.

¹ Sotelo, C.I. 1997. Informe de comisión de campo Macizo de Floresta. Ingeominas, Informe preliminar (Inédito).

² Jiménez, D.M. 2000. Catálogo de las Unidades Litoestratigráficas de Colombia: Filitas y Esquistos de Busbanzá. Ingeominas, Informe Interno: 17 p.

resources (*e.g.*, Alvarado and Sarmiento, 1944), stratigraphical (*e.g.*, Cediél, 1969; Mojica and Villarroel, 1984), paleogeographical (*e.g.*, Barrett, 1983, 1988; Janvier and Villarroel, 1998, 2000; Berry *et al.*, 2000; Moreno-Sánchez, 2004), and structural (*e.g.*, Kammer and Sánchez, 2006) interest.

The Colombian Devonian fauna and flora has been scarcely studied (*e.g.*, Olsson and Caster, 1937; Caster, 1939; Royo y Gómez, 1942; Barrett, 1983, 1988; Berry *et al.*, 2000; Janvier and Villarroel, 2000; Moreno-Sánchez, 2004) and is characterized by the occurrence of abundant and diverse plants, invertebrates, and vertebrates. However, although fossils are widely distributed in the Floresta Massif, their preservation is often poor (Morzadec *et al.*, 2015). In addition, studies on the Floresta Formation have reported thicknesses that vary between 300 m to the north and 700 m to the south of the Floresta Massif (Botero, 1950; Mojica and Villarroel, 1984; Ingeominas, 2003; Mozardec *et al.*, 2015), which may be evidence of an asymmetric basin that deepens southwards and is affected by the structural complexity of the area. The influence of deformation on the thickness variations of the Floresta Formation has, however, not been tested yet; neither has how this deformation is different from overlying units. To tackle this problem, clay minerals were used. Clay minerals are abundant in the sedimentary record of the Floresta Formation, and can be useful paleoenvironmental indicators (*e.g.*, Jacobs and Hays, 1972; Singer, 1984; Środoń, 2002). Illite crystallinity (IC) indexes in specific samples of the Floresta Formation were analyzed to estimate their burial diagenesis conditions, and clay mineralogical analyses were conducted to determine their maximum burial depths and temperatures. This paper therefore seeks to make contributions on the sedimentology, paleoenvironmental interpretations, post-sedimentary history, and structures of the Floresta Formation, by means of information from field campaigns, geological mapping, laboratory analyses, and structural cross sections.

2. Geological setting

2.1. Floresta Massif

The Floresta Massif, oriented in the NE-SW-direction, is situated in the axial zone of the Eastern Cordillera of the Colombian Andes (Figs. 1, 2). The

Floresta Massif is the core of a broad anticline as defined by Cretaceous and Cenozoic rocks, delimited on its flanks by two regional reverse faults: the Boyacá Fault to the west and the Soapaga Fault to the east (Fig. 2). Kammer and Sánchez (2006) reported Early Jurassic rift structures associated with the Soapaga and Boyacá faults that were inverted since Paleogene time (*e.g.*, Saylor *et al.*, 2012; Bayona *et al.*, 2013). The core of this ancient massif consists of metamorphic rocks of the Busbanzá's Phyllites and Schists Formation of Cambrian-Ordovician age (or older), which were affected by a first event of regional metamorphism of medium P-T conditions that was overprinted by a thermal event (contact metamorphism) during the Ordovician related to the emplacement of a series of syn-tectonic granitic intrusions (Otengá Stock), which developed cordierite hornfels (Manosalva-Sánchez *et al.*, 2017).

The crystalline core of the Floresta Massif is overlain by a sedimentary sequence of Emsian to Famennian age (Moreno-Sánchez *et al.*, 2020). This sedimentary sequence was divided into three units (Botero, 1950): the El Tibet Formation of Emsian age (Grösser and Prössl, 1994), the Floresta Formation of Late Emsian to Givetian age (Morzadec *et al.*, 2015), and the Cuche Formation of Late Devonian age at the top (Mojica and Villarroel, 1984; Janvier and Villarroel, 2000). The Cuche Formation is unconformably overlain by Jurassic sedimentary rocks of the Girón Formation (Bayona *et al.*, 2020).

2.2. Devonian successions

The first report on the occurrence of Devonian rocks in Colombia can be attributed to Axel A. Olsson and Parke A. Dickey, geologists that worked during the 1930s for the petroleum industry and that made a significant discovery in the Floresta Massif (Moreno-Sánchez *et al.*, 2020). They collected fossils in these rocks north of the Floresta town in the Boyacá Department, which were further examined by Caster (1939) and McNair (1940). This term Floresta Series was initially proposed by Olsson and Caster (1937) for a sedimentary sequence between the metamorphic units and the Girón Group. Later, Botero (1950) introduced the term Floresta Formation for three sedimentary sets: a lower set of conglomeratic sandstones with a maximum thickness of ~30 m; a middle set of yellowish to violet claystones with a maximum thickness of ~530 m; and an upper set of sandstones

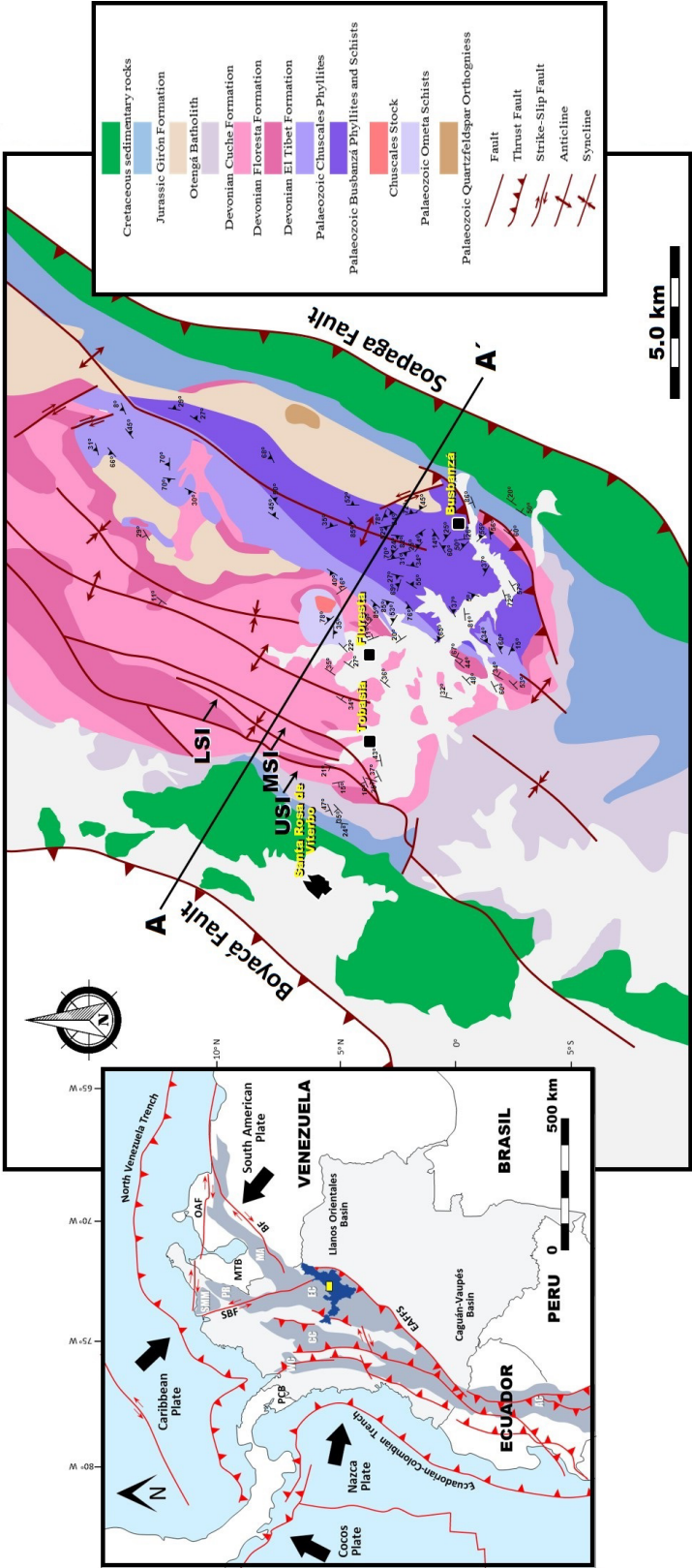


FIG. 2. Left figure: Major tectonic and structural features of the NW South American margin (modified from Colmenares and Zoback, 2003). SMM, Santa Marta Massif; PR, Perijá Range; MA, Merida Andes; AC, Andes Cordillera; OAF, Oca-Ancón Fault; SBF, Santa Marta-Bucaramanga Fault; BF, Boconó Fault; MTB, Maracaibo Triangular Block; PCB, Panamá-Chocó Block; EC, Eastern Cordillera; CC, Central Cordillera; WC, Western Cordillera; and EAFFS, Eastern Andean Front Fault System. The Department of Boyacá is shown in blue and the study area as a yellow rectangle. Right figure: Generalized geological map of the Floresta Massif in the Eastern Cordillera of the Colombian Andes (adapted and modified from Jiménez, 2000)², showing the location of studied stratigraphic sections and the A-A' structural cross-section. LSI: The Lower Stratigraphic Interval in the Teneria and Monticelo villages; MSI: The Middle Stratigraphic Interval in the Tablón village; USI: The Upper Stratigraphic Interval in the Chuscales and Tobasia villages.

with a maximum thickness of ~150 m. Mojica and Villarroel (1984) divided the Floresta Series into three units: the El Tibet Formation at the base, the Floresta Formation in the middle, and the Cucho Formation at the top. They used the term Floresta Formation for the succession that discordantly overlies metamorphic rocks and concordantly overlies the El Tibet Formation, and transitionally underlies the Cucho Formation. Its upper limit with the Cucho Formation is locally discordant (Botero, 1950). The Early Devonian El Tibet Formation, which unconformably overlies Ordovician granites and Early Paleozoic metamorphic rocks, consists of a fining-upward sequence (up to 420 m thick) of conglomerates, sandstones, and interbedded gray-colored shales, signaling the Devonian marine transgression onset (Morzadec *et al.*, 2015). Cedié (1969) initially considered the El Tibet Formation as part of the Floresta Formation, but Mojica and Villarroel (1984) later recognized it as a different formation.

The Early to Middle Devonian Floresta Formation consists of a 600-m-thick monotonous succession of highly fossiliferous (brachiopods, bryozoans, crinoids, trilobites, ostracods, corals, mollusks, and vertebrates) siltstones. Barrett (1983) first interpreted the depositional environment of the Floresta Formation as a typical beach with or without influence of waves. Later, it was suggested that the Floresta Formation was deposited in a transgressional and regressional epicontinental marine environment at the edge of the Paleo-Tethys Ocean (Janvier and Villarroel, 1998). Barrett (1988) considered that the Floresta Formation was deposited in a shallow, epicontinental sea opened towards the north during the Emsian-Frasnian time; however, according to Giroud-López (2014), the lower and upper parts of this formation were deposited under coral reef and deltaic environments, respectively. The Late Devonian Cucho Formation, is characterized by a 550-m-thick sequence of alternating pinkish to greenish, sandy, silty, and argillaceous levels yielding plant fragments, vertebrate remains, mollusks and ostracods (Morzadec *et al.*, 2015), deposited under both continental and transitional marine environments (e.g., Janvier and Villarroel, 2000; Moreno-Sánchez, 2004). A regressive event is also recorded in the Cucho Formation (Morzadec *et al.*, 2015). Several ages have been attributed to the Floresta Formation, which can be considered as Emsian±Eifelian. This age interval has been established on the basis of its fossiliferous content, essentially brachiopods

(e.g., Caster, 1939; McNair, 1940; Royo y Gómez, 1942; Barrett 1985, 1988). However, Morzadec *et al.* (2015), based on the study of Devonian Colombian trilobites and inarticulate brachiopods, recognized two biostratigraphic levels in the Floresta Formation characterized by their markedly different lithofacies and faunal contents: the first level (lower part) of Late Emsian age (Early Devonian) and the second level (upper part) of Givetian age (Middle Devonian).

Here we suggest a subdivision of the Floresta Formation into three intervals, derived from a thorough analysis encompassing major structural features, stratigraphy, variations in fossil content, lithology, and vein networks, as well as correlation criteria. This proposed division enhances our comprehension of the intricate geological history preserved in the Floresta Massif throughout the Early to Middle Devonian. Beyond serving as a noteworthy discovery in our research, this subdivision lays the groundwork for future investigations into the tectonic evolution, sedimentary processes, and environmental changes within the region.

The actual configuration of the Floresta Massif and the Eastern Cordillera is the result of uplift and exhumation induced by displacement along most external inverted faults of the Eastern Cordillera Foothills (e.g., Mora *et al.*, 2013; Reyes-Harker *et al.*, 2015). These recent events affected the sedimentary rocks of interest in the present study exposing them to differential deformation and ultimately affected by denudational processes that have contributed to the modeling of the current landscape.

3. Fieldwork and laboratory methods

3.1. Fieldwork and structural analysis

Several fieldwork campaigns were performed by a team of students and researchers of the School of Geology of the Universidad Industrial de Santander in the Floresta Massif. Based on stratigraphic features recognized in the field and already mentioned in previous works (e.g., Botero, 1950; Mojica and Villarroel, 1984; Ingeominas, 2003), such as the identification of sites where the lower and upper contacts of the Floresta Formation with the El Tibet and Cucho formations can be observed, key stratigraphic sections were measured using the Jacob's staff method. This was complemented by lithology and stratigraphic descriptions, structural data collection, rocks sampling for clay mineral analyses, and collection of fossil specimens from

several outcrops in the Floresta Formation. Drawing upon the collected stratigraphic and structural data, a comprehensive structural cross-section was developed for three primary purposes: (1) recognizing the geometry of existing structures, (2) deriving accurate stratigraphic thicknesses of the Floresta Formation by identifying stratigraphic repetitions within the sequence, and (3) gaining insights into the temporal aspects of deformation events. The structural cross-section was made by integrating the main rock outcrops and the field measured sections of the Floresta Formation, crossing as many structures as possible.

3.2. Analytical methods

The clay minerals present in the sedimentary rocks of the Floresta Formation underwent a comprehensive analysis to examine both their mineralogical compositions and their illite crystallinity (IC) indexes. Additionally, the prevailing conditions of burial diagenesis were determined. Mineralogical characterization in bulk rock samples was carried out from X-ray powder diffraction (XRPD) patterns, which were obtained using a BRUKER D8 ADVANCE X-ray diffractometer operating in Da Vinci geometry and equipped with an X-ray tube (Cu-K α 1 radiation: $\lambda=1.5406$ Å), a 1-dimensional LynxEye detector (aperture angle of 2.93°), a divergent slit of 0.6 mm, two Soller axials (primary and secondary) of 2.5° and a nickel filter. All samples were ground in an agate mortar to a particle size of less than 50 μ m and then mounted on a sample holder of polymethylmethacrylate (PMMA) using the filling front technique prior to XRPD analysis. Data collection was carried out at 40 kV and 30 mA in the 2 θ range of 3.5-70°, with a step size of 0.01526° (2 θ) and a counting time of 1 s/step. Phase identification was performed using the crystallographic database Powder Diffraction File (PDF-2) from the International Centre for Diffraction Data (ICDD) and the Crystallographica Search-Match program. The unit-cell constants, atomic positions, factors of peak broadening and phase concentrations were refined and calculated using the MDI RIQAS program based on the Rietveld method. The IC index was calculated according to the full-width half maximum (FWHM) of the main illite (001) peak. IC calibrated using the Crystallinity Index Standard (CIS) scale (Warr and Rice, 1994) through the equation $IC(CIS)=0.93IC+0.09$ (Dellisanti *et al.*, 2008). Morphological features and elemental composition of the mineral phases were

examined by Scanning Electron Microscopy (SEM), using secondary electron (SE) imaging and Energy Dispersive X-ray Spectroscopy (EDS), respectively, on a FEI QUANTA 650 FEG-ESEM, under the following analytical conditions: magnification=100-20,000x, Working Distance=9.0-11.0 mm, High Vacuum Mode=20 kV, signal=Backscattered electron in ZCONT mode, detector=Backscattered electron detector, EDS Detector EDAX APOLLO X with a resolution of 126.1 eV (in. Mn K α). The rock-forming minerals of the Floresta Formation mudstones were investigated through semi-quantitative EDS analysis, aiming to analyze their composition and characterize the mineralogy of the samples.

4. Results

In this section we present a detailed analysis of the major structural features and mapped stratigraphic units identified in the newly developed geological map. The observations and interpretations are based on comprehensive fieldwork and structural analyses conducted in the study area. Our focus lies on elucidating the intricate geological history recorded in the Floresta Formation, particularly in the Lower, Middle, and Upper Stratigraphic Intervals (LSI, MSI, and USI). Through a systematic examination of the stratigraphic sequences and their most distinctive structural elements, we aim to provide a nuanced understanding of the tectonic and sedimentary processes shaping the Floresta Massif. Additionally, we emphasize the significance of tight anticlines recognized in the LSI interval, discussing criteria for identifying normal and overturned beds. The following discussion unfolds the complexities inherent in the structural evolution of the study area, shedding light on the interplay between tectonics, sedimentation, and diagenesis during the Middle Devonian period.

4.1. Structural analysis

In the study area, a repetition of the sequence is observed based on stratigraphic intervals and the presence of ferruginous veins and fossil content. We interpret this as overturned folds within the Floresta Formation (Figs. 2, 3). The structural style and deformation in the study area is illustrated in the structural cross sections of figure 3, which allows visualizing the differences between the folds developed in the Early to Middle Devonian Floresta Formation and those observed in the underlying El

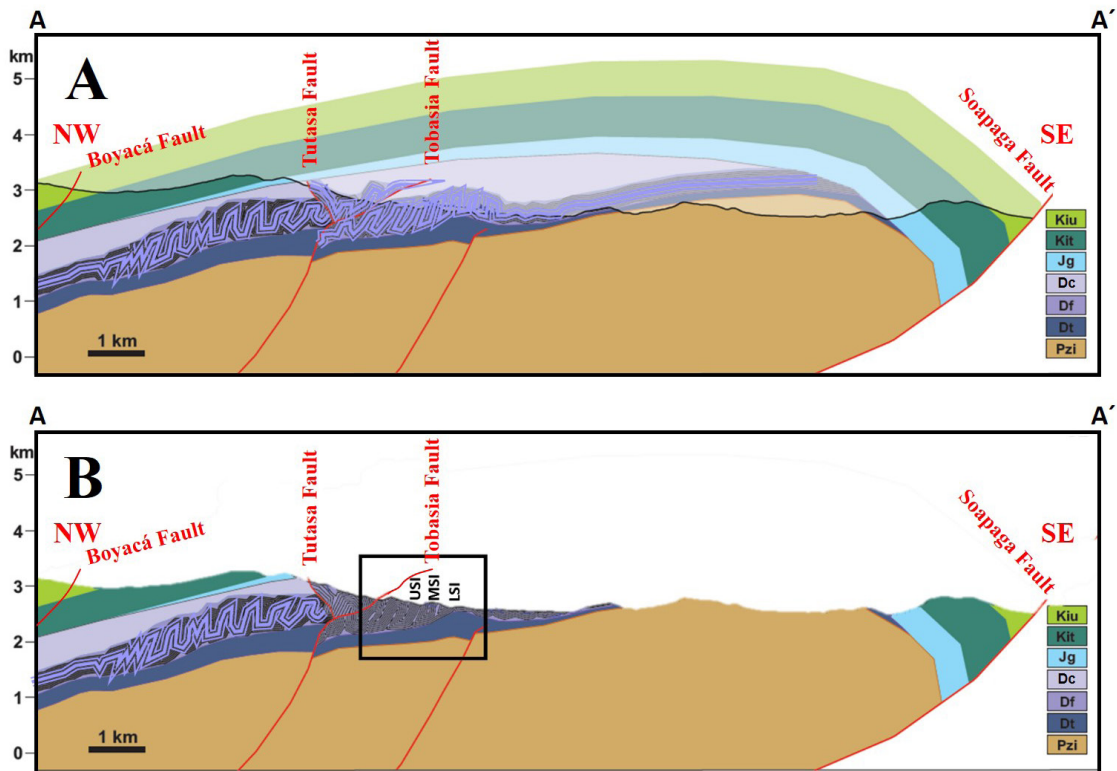


FIG. 3. Geological cross-sections A-A', **A.** without erosion and **B.** with erosion, of the Floresta Anticline limited by the Boyacá and Soapaga faults, showing the following geological units in lithostratigraphic order from oldest to youngest: Metamorphosed Floresta rocks (Pzi), El Tibet (Dt), Floresta (Df), Cuche (Dc), Girón (Jg), Tibasosa (Kit), and Une (Kiu) formations. See figure 2 for ubication. In (B), **LSI**: The Lower Stratigraphic Interval; **MSI**: The Middle Stratigraphic Interval; and **USI**: The Upper Stratigraphic Interval. Black square refers to a detailed excerpt shown in figure 4, highlighting the three stratigraphic intervals.

Tibet Formation and the overlying Cuche Formation, south of the western flank of the Floresta Massif (Figs. 2-4). Furthermore, it is worth noting that the younger strata of the Jurassic Girón Formation and the Early Cretaceous Tibasosa Formation are not affected by such intense deformation, corresponding solely to smooth and broad folds (Fig. 3). The structural information collected shows that the folds in the underlying (El Tibet) and overlying (Cuche) strata are smooth and with low deformation. The Floresta Formation, on the other hand, has narrow and tight folds, with beds that pinch out eastwards (Figs. 3, 4).

The largest folds are called first-order folds, whereas smaller associated folds are second- and higher-order folds, also called parasitic folds. First-order folds can be of any size, but they have to be observable at a map

scale, while second or higher-order folds are only visible at the outcrop scale (e.g., Fossen, 2010; Bürg, 2017)³. This hierarchical approach to fold classification adds a layer of complexity that enriches our interpretation of the tectonic and structural history of the Floresta Formation. Parasitic folds, although less evident on a regional scale, provide valuable information about more local geological events, thus contributing to a more detailed understanding of deformation in the study area.

4.2. Stratigraphy and rock description

4.2.1. Lower Stratigraphic Interval

The Lower Stratigraphic Interval (LSI) has a maximum thickness of ~65 m (Fig. 5). The LSI consists of intercalations of white-reddish silty and

³ Bürg, J.P. 2017. Folding. ETH Zürich and Universität Zürich. <http://www.files.ethz.ch/structuralgeology/JPB/files/English/7folding.pdf>. (Last visit 15/01/2020).

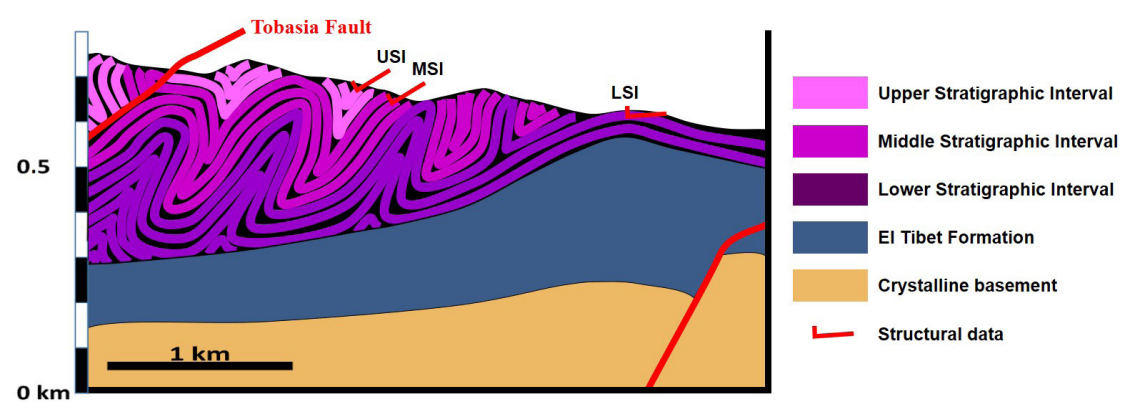


FIG. 4. Cross section in detail showing outcrops of the Lower (LSI), Middle (MSI), and Upper (USI) stratigraphic intervals in the Floresta formation. Note the complex deformation and the thickness of each stratigraphic interval. See black rectangle in figure 3B for lithology and location.

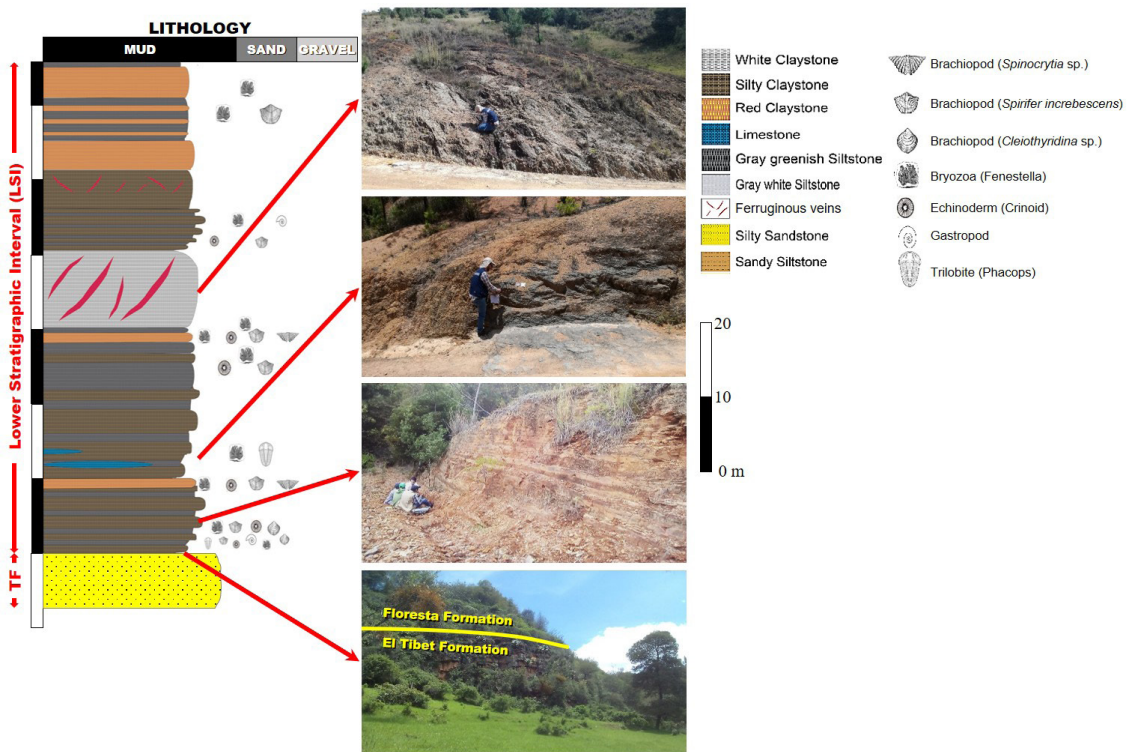


FIG. 5. Stratigraphic column of the LSI. Note in the second picture (top-down) the lenses of limestone. The fossil content is high at the bottom and decreases up-section. The veins of iron oxides are null in the lower half. TF: El Tibet Formation. See figure 2 for column location.

greenish silty claystones with planar and discontinuous lamination. To the base of the LSI, which shows a conformable contact with the El Tibet Formation,

there are well-selected and cemented medium-grained quartz-sandstones light grey, yellowish, and white in color, as well as ocher and greenish gray siltstones,

fossiliferous claystones, and lenses of fossiliferous limestones. Close to the upper contact with the rocks of the middle stratigraphic interval (MSI), a network of irregular veinlets of ferruginous oxides, red and orange in color, occur. The fossil record of the LSI includes brachiopods, bryozoans, gastropods, trilobites, corals, bivalves, ostracods, and rare plant remains. Mainly trilobites of the species *Phacops sp.*, *Viaphacops cristata*, *Synphoria stemmata* and *Dalmanites sp.* are found in this interval. This fossil record has been interpreted in previous studies as characteristic of the Middle Devonian in Colombia (e.g., Caster, 1939; Royo y Gómez, 1942; Morales, 1965). All these evidences allow us to recognize the repetition of the LSI in other places.

4.2.2. Middle Stratigraphic Interval

The Middle Stratigraphic Interval (MSI) has a maximum thickness of ~90 m (Fig. 6). It mainly consists of intercalations of red claystones and greenish gray siltstones with planar, undulated and discontinuous lamination, with intercalations of gray and black siltstones at the top. The main feature of this interval, and the one

that allows to recognize the repetition of the sequence in other places, is the presence of abundant ferruginous vein networks (stockworks). The veins are ~0.5 to 2 cm wide and can reach up to several meters in length. Most of the veins are partially to entirely oxidized, which has resulted in the formation of orange-brown limonites ($\text{FeO} \cdot \text{OH} \cdot n\text{H}_2\text{O}$). Hematite is the most abundant iron oxide and occurs either as disseminated coarse grains or as aggregates closely intergrown with quartz. It shows a red to reddish brown color, with local yellowish oxidation patina and locally earthy submetallic luster, and it can show empty chambers of irregular shape. There are three types of veinlets: (1) veins that do not have a preferential orientation and cut each other chaotically and in great quantity; (2) those showing two preferential orientations (azimuths 0-10° and 80-90°) and the veinlets cut each other at right angles; and (3) those parallel and with a preferential orientation.

4.2.3. Upper Stratigraphic Interval

The Upper Stratigraphic Interval (USI) has a maximum thickness of ~55 m (Fig. 7). It is mainly



FIG. 6. Stratigraphic column of the MSI. In the middle picture, the iron oxide veins have two preferred orientations. The fossil content is typically poor. See figure 2 for column location.

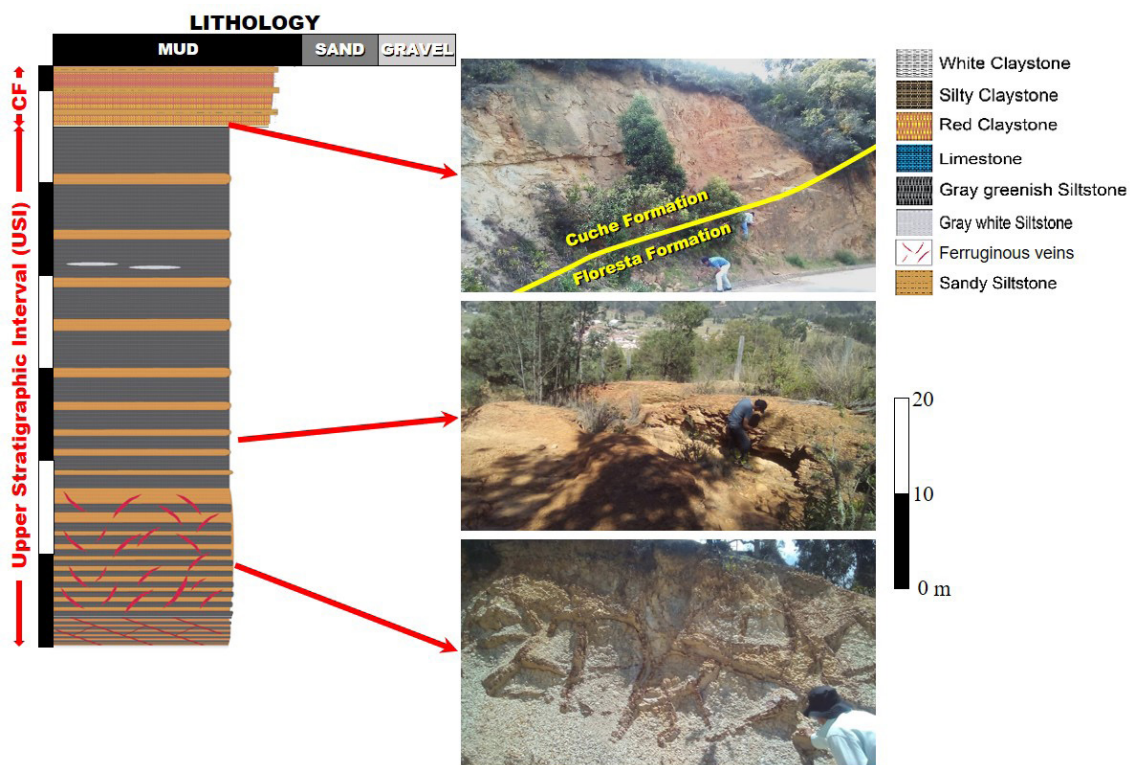


FIG. 7. Stratigraphic column of the USI. In the lower picture the iron oxide veins do not have a preferred orientation. CF: Cuche Formation. See figure 2 for column location.

characterized by intercalations of gray and mottled siltstones and red claystones with discontinuous lamination. Its top shows a transitional contact with the Cuche Formation. The fossil record and ferruginous vein networks at top of the sequence are very scarce to null, facilitating the recognition of any repetition of this interval in other places. Although, Giroud-López (2014) reports for this part of the Floresta Formation thin lenticular layers of siltstone with presence of cephalopods and unarticulated brachiopods, as well as of plant remains, no evidence of these species was found in the section studied.

4.2.4. Composite stratigraphic column

As a result of this work, it was possible to recognize three intervals for the Floresta Formation, which are briefly described from base to top (Fig. 8). We integrate these intervals into one single column, ~210 m-thick (Fig. 8). At the bottom, the contact between the Lower Stratigraphic Interval (LSI) and the El Tibet Formation is observed in the Monticelo

village (Teneria sector), the Middle Stratigraphic Interval (MSI) can be well described in the Tablón sector, and the transitional contact between the Upper Stratigraphic Interval (USI) and the Cuche Formation in the Santa Rosa sector (Fig. 2). The main differences between the three intervals are the presence of fossiliferous levels (LSI, Figs. 5, 8), ferruginous veins (MSI, Figs. 6, 8) and silty beds (USI, Figs. 7, 8). The principal sedimentary structures in the three intervals are planar lamination, and undulated and discontinuous lamination.

Below, we discuss some relevant aspects upon which the subdivision of the Floresta Formation into the three intervals mentioned above is based. The LSI's rich fossil content suggests favorable conditions for diverse marine life, possibly indicating a shallower and stable marine environment. The scarcity of fossils in the USI may be attributed to rapidly changing environmental conditions or sedimentary dynamics. The MSI's abundance of ferruginous veins indicates significant fluid-rock interaction, possibly related to

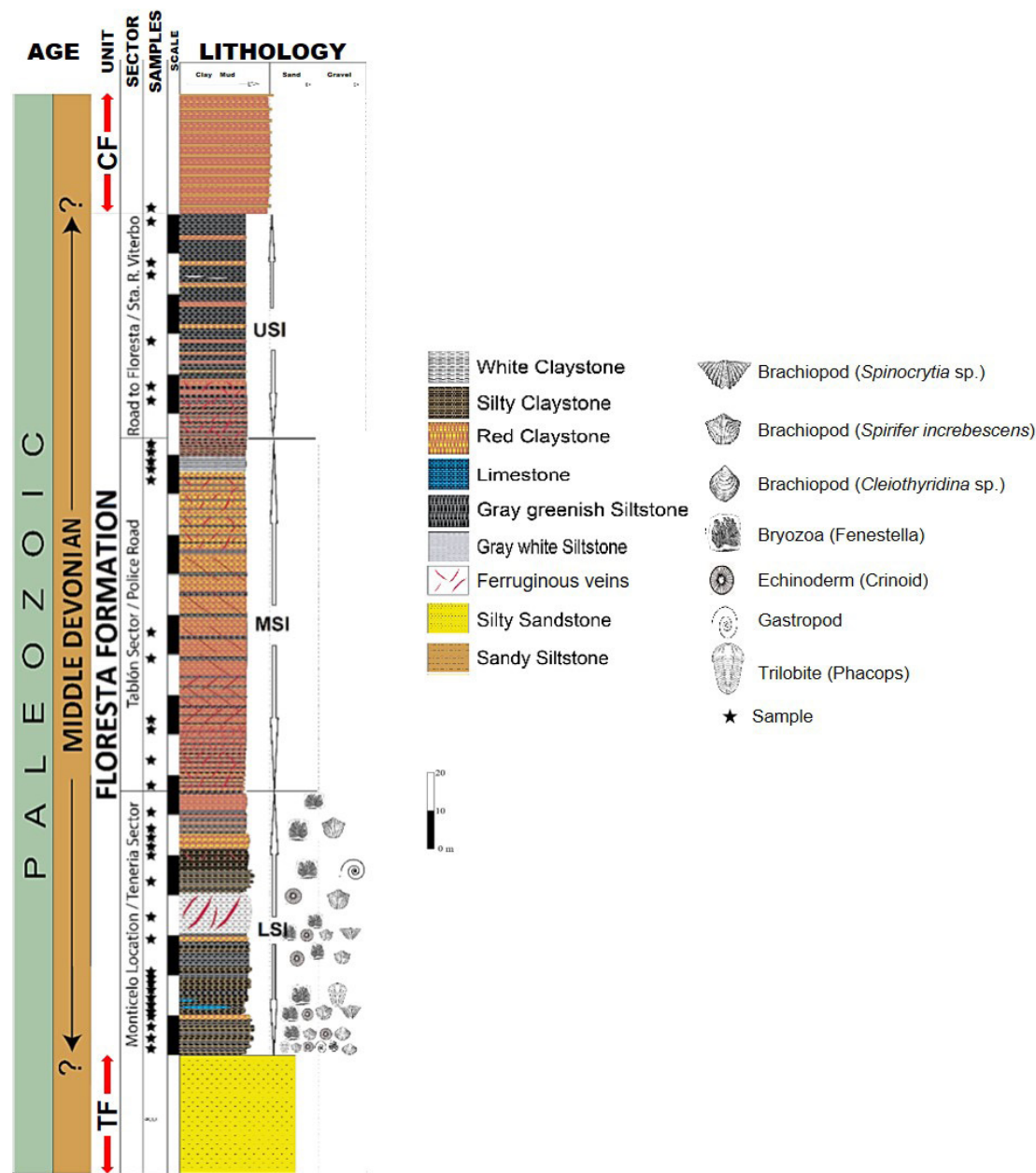


FIG. 8. Lithostratigraphy of the Floresta Formation in the study area, made after collating information from the field sections seen in figures 5-7. LSI: Lower Stratigraphic Interval; MSI: Middle Stratigraphic Interval; USI: Upper Stratigraphic Interval.

hydrothermal activity. This feature is absent (or less pronounced) in the other intervals, suggesting temporal variations in fluid flow. The lithological differences between the intervals, such as the dominance of intercalations of claystones and siltstones in the LSI, claystones in the MSI, and claystones and siltstones

in the USI, reflect changes in sediment sources, transport, and deposition. These variations could be linked to shifts in sea level or tectonic events. The nature of the contacts, particularly the conformable contact with the El Tibet Formation in the LSI and the transitional contact with the Cuche Formation in

the USI, provides insights into the depositional history and possible environmental changes. The significant development of ferruginous veins in the MSI suggests a distinct diagenetic history compared to the other two intervals.

The observed variations in fossil content, lithology, and vein networks among the three intervals of the Floresta Formation indicate varying geological and environmental conditions during the Middle Devonian. The following evidence was used here to systematically characterize the three intervals of the Floresta Formation: (1) lithology; (2) fossil content; (3) vein networks; (4) sedimentary structures; (5) transitional contacts; and (6) thickness variations. The lithological characteristics of each interval were compared; for example, the presence of well-selected and cemented medium-grain quartz-sandstones in the LSI, the dominance of ferruginous veins in the MSI, and the transitional contact with the Cuche Formation in the USI.

Fossil assemblages, particularly the presence of trilobites like *Viaphacops cristata* and *Synphoria stemmata* were also used for correlation. In particular, the presence of two distinct trilobite associations in the Floresta Formation holds significance in understanding the paleogeographic evolution of the region during the Devonian (Morzadec *et al.*, 2015). The trilobite association aligns with the North Eastern Americas Realm, primarily recognized from the Appalachian Basin to the Gaspé region of Canada (Lespérance and Sheehan, 1988). In the Appalachian Basin, the genera *Viaphacops cristata* (Eldredge, 1973) and *Synphoria stemmata* (Lespérance and Bourque, 1971) represent typical Early Devonian fauna. In our study area, the trilobite association can be less diverse, although it maintains some North American affinities of the Middle Devonian fauna of this region (Morzadec *et al.*, 2015). In general, there are great similarities between the fauna associations of the Floresta Formation and those of North America, something that was supported by Morales (1965) mainly based on brachiopod associations. Carvalho and Moody (2000) established close affinities in Colombian faunal content with the Middle Devonian of Venezuela, which suggests that Colombia and Venezuela were geographically close and shared strong faunal similarities with Europe and North America (Cooper, 1982). Based on the information presented above, particularly in the trilobite and inarticulate fauna associations, it can be inferred

that two distinct biostratigraphic levels exist in the Floresta Formation. The trilobite associations suggest an Early Devonian age for the lower part of the Floresta Formation, while trilobite and brachiopod associations indicate a Middle Devonian age for its upper part. Therefore, the content of specific fossils in each interval can be used as biostratigraphic markers.

The distinctive ferruginous vein networks in the MSI, including their orientation and distribution, is another useful correlation criterion; so is the nature of sedimentary structures, such as planar lamination, undulated lamination, and discontinuous lamination. Variations in sedimentary structures can indicate changes in depositional conditions. Any differences in the thickness of the intervals, especially in comparison to previous studies, can be used as a correlational factor as well. Significant variations in thickness might indicate changes in sedimentation and/or tectonic activity.

4.3. Clay mineralogy

The analysis of clay minerals remains a major challenge due to their various chemical compositions, preferred orientation, structural disorder and great structural diversity (e.g., Środoń, 2002; Bergaya and Lagaly, 2006). Table 1 summarizes the results of the analyzed samples by XRPD and SEM/EDS methods, which are described below. A very homogeneous composition in the clay mineralogy can be observed for the three intervals, with illite and kaolinite as major minerals and quartz, chlorite and oxides as accessory minerals.

4.3.1. X-ray powder diffraction analysis (XRPD)

Figure 9 shows the XRPD patterns of the analyzed rock samples. Clay minerals like illite, with minor kaolinite, chlorite, smectite and montmorillonite are identified, together with Fe-oxides and quartz (Table 1). Illite and kaolinite are ubiquitous throughout the Floresta Formation sedimentary sequence. The intensity peaks of illite are higher in figures 9B-9C. Although the XRPD patterns shown in figure 9 do not clearly reveal the presence of interstratified (or mixed layer) clay minerals (I/S) such as illite/smectite or kaolinite/smectite, they are sometimes observed in the SEM images (Fig. 10). A regularly interstratified illite-smectite is present around $2\theta=7-8^\circ$ (Fig. 9). In figure 9D, a peak at $2\theta=6.275^\circ$ is enclosed in a yellow rectangle, which can be attributed to the presence of smectite, chlorite,

TABLE 1. RESULTS OF THE XRPD AND SEM/EDS ANALYSES OF CLAY MINERALS.

Sample	Level	Lithology	XRPD		SEM	
			Clay minerals	Other minerals	Clay minerals	Other minerals
FLO-1	USI	Red claystone	Chl, Ill	Qtz, Ox	Ill, Kao	Ox
FLO-2	USI	White claystone			Ill	Qtz, Zrc
FLO-3	USI	Greenish gray siltstone	Ill, Kao	Qtz	Ill, Kao	Qtz, Ox
FLO-4	MSI	Black siltstone	Ill	Qtz	Ill, Kao	Qtz, Ox
FLO-5	MSI	Light gray siltstone			Ill	Ox
FLO-6	MSI	Vein	Ill, Kao	Qtz, Ox	Ill	Qtz, Ox
FLO-7	MSI	Greenish gray siltstone			Ill, Kao	Qtz, Ox
FLO-8	MSI	Vein			Ill, Kao	Qtz, Ox
FLO-9	MSI	Greenish gray siltstone	Ill, Kao	Qtz	Ill, Kao	Qtz, Ox
FLO-10	LSI	Red claystone			Chl, Ill	Qtz, Ox
FLO-11	LSI	Silty claystone			Ill, Kao	Qtz, Ox
FLO-12	LSI	Vein	Ill, Kao	Qtz, Ox	Ill, Kao	Qtz, Ox
FLO-13	LSI	Greenish gray siltstone			Chl, Ill, Kao	Qtz, Ox
FLO-14	LSI	Greenish gray siltstone	Ill, Kao	Qtz	-	-
FLO-15	LSI	Limestone			Ill, Chl	Cal
FLO-16	LSI	Red claystone	Ill, Kao	Qtz, Ox	Ill, Kao	Qtz, Ox
FLO-17	LSI	Silty claystone			Ill, Kao	Qtz, Zrc
FLO-18	LSI	Red claystone	Chl, Ill, Kao	Qtz	Kao, Ill, Chl	Qtz, Ox

LSI: Lower Stratigraphic Interval; MSI: Middle Stratigraphic Interval; USI: Upper Stratigraphic Interval. Ill: illite; Kao: kaolinite; Chl: chlorite; Qtz: quartz; Cal: calcite; Ox: oxides; Zrc: zircon.

and/or montmorillonite. This peak does not move after applying ethylene glycol, suggesting the presence of chlorite unlike figure 9A. Minor vermiculite (results not shown) seems to be present as well. Vermiculite is a fast-forming and unstable intermediate mineral phase, derived from muscovite which can then be transformed to other minerals, depending on the environment (e.g., Velde and Meunier, 2008).

Several clay mineral assemblages in the Floresta Formation can be recognized: illite-kaolinite, chlorite-illite-kaolinite, smectite-chlorite-montmorillonite, and illite-chlorite-I/S. However, the quantification of clay minerals remains complex relative to the quantification of other minerals, taking into account that clay minerals have structures characterized by various polytypes and types of defects, and can vary in chemical composition and in terms of effects their preferred orientation (e.g., Lippmann, 1970; Środoń *et al*, 2001; Środoń, 2002; Viennet *et al.*, 2016).

4.3.2. SEM analysis

There is a good correlation between the degree of alignment of illite and/or kaolinite and the elastic anisotropy of the rock, as well as between the mineralogy and the strength and stiffness of the rock (e.g., Voltolini *et al.*, 2009). Clay minerals (mainly illite and/or kaolinite floccules) can develop a “cardhouse” structure of individual edge-face- or edge-edge-oriented flakes, something seen in figure 10 and reported by other authors elsewhere (e.g., Bennett *et al.*, 1991; Slatt and O’Brien, 2011). The SEM images allow recognizing several pore types (Fig. 10), which according to Slatt and O’Brien (2011), include interparticle pores produced by flocculation, intraparticle pores from fossils, intraparticle pores within mineral grains and microchannels and microfractures.

The analyzed sedimentary rocks of the Floresta Formation show a loose texture and are poorly

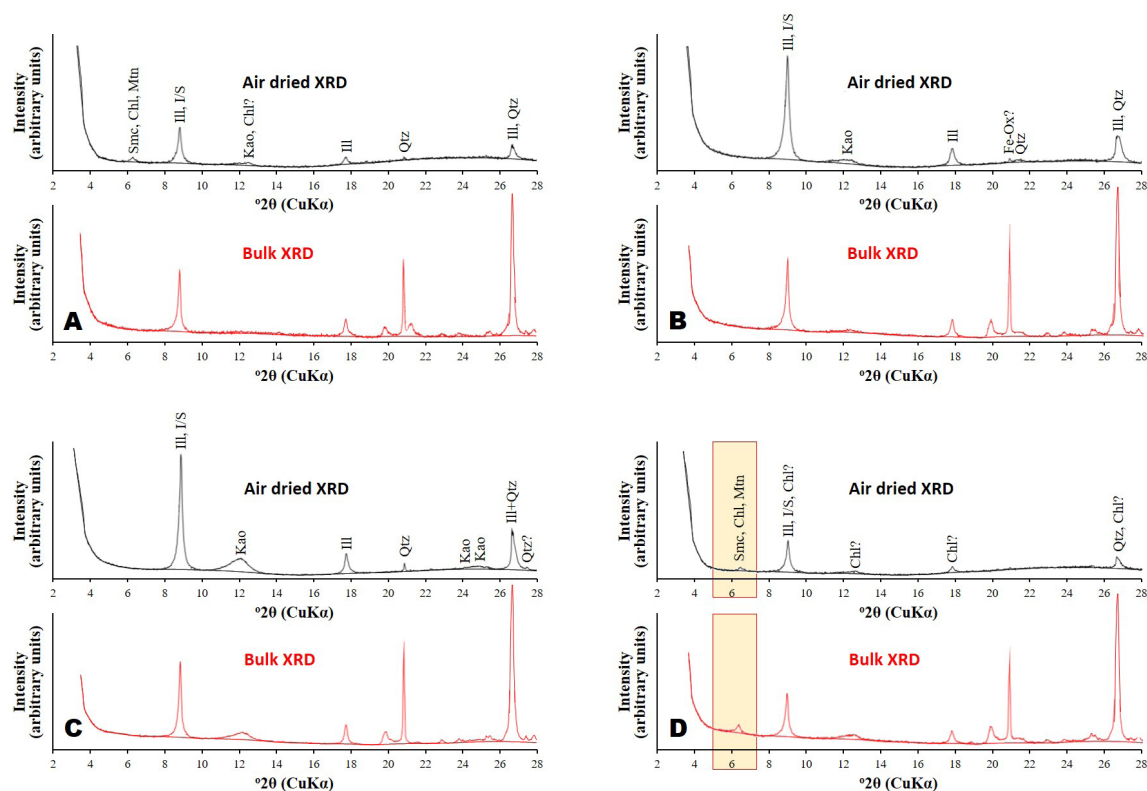


FIG. 9. XRPD patterns of representative rocks of the Floresta Formation. Air-dried XRPD patterns (or drying at room temperature) are shown in black color, whereas bulk XRPD patterns (sample in powder size) are shown in red color. The samples are in order of stratigraphic intervals, bottom-up: **A.** Sample FLO-18, LSI; **B.** Sample FLO-10, LSI; **C.** Sample FLO-4, MSI; and **D.** Sample FLO-1, USI. See table 1 for mineral compositions and lithologies.

cemented. Clay minerals are identified as the main mineral phases. The microstructure of the analyzed rocks shows the characteristic flaky and blocky morphologies of clays and quartz, respectively, and the occurrence of isolated pores within the rock matrix (Fig. 10A-F). Figures 10A and 10D illustrate microchannel pathways, which allow the migration or flow of fluids. Note the occurrence of iron oxides and quartz, which show spherical and blocky morphologies, respectively. In the samples analyzed (Fig. 10), clay minerals usually occur as particles that exhibit irregular and rounded outlines, indicating a detrital origin (e.g., Jouanneau and Latouche, 1981; Hong *et al.*, 2012; Turner and Huggett, 2019; Virolle *et al.*, 2019). In general, clay minerals show a flaky morphology with a bent shape and relatively smooth basal (001) planes, and their lateral surfaces are very uneven, with irregular outlines or ragged edges. According to Hong *et al.*

(2012), the lateral dimensions of the clay flakes are poorly defined, with particularly thin plates and well-developed fissures, suggesting an origin in detrital clasts. Both illite and illite/smectite may also occur with a ragged or platy morphology, developing sometimes lath-shaped overgrowths (e.g., Pollastro, 1985; Inoue *et al.*, 1988; Hugget, 1995).

A detrital origin for most of the illite and illite/smectite is inferred from the ragged appearance of the flat (illite) to undulose (illite/smectite) platelets (e.g., Hugget, 1995; Slatt and O'Brien, 2011). Kaolinite appears as detrital grains, mainly as a dispersed mineral phase filling pores between clastic grains or constituting the planar structure of the rock. It shows a massive appearance that is clearly different from those of both illite and illite/smectite (Keller *et al.*, 1986). Chlorite, like illite and illite/smectite, is arranged in elongated leafy forms (Fig. 10) with a certain preferential orientation, with fiber-type textures

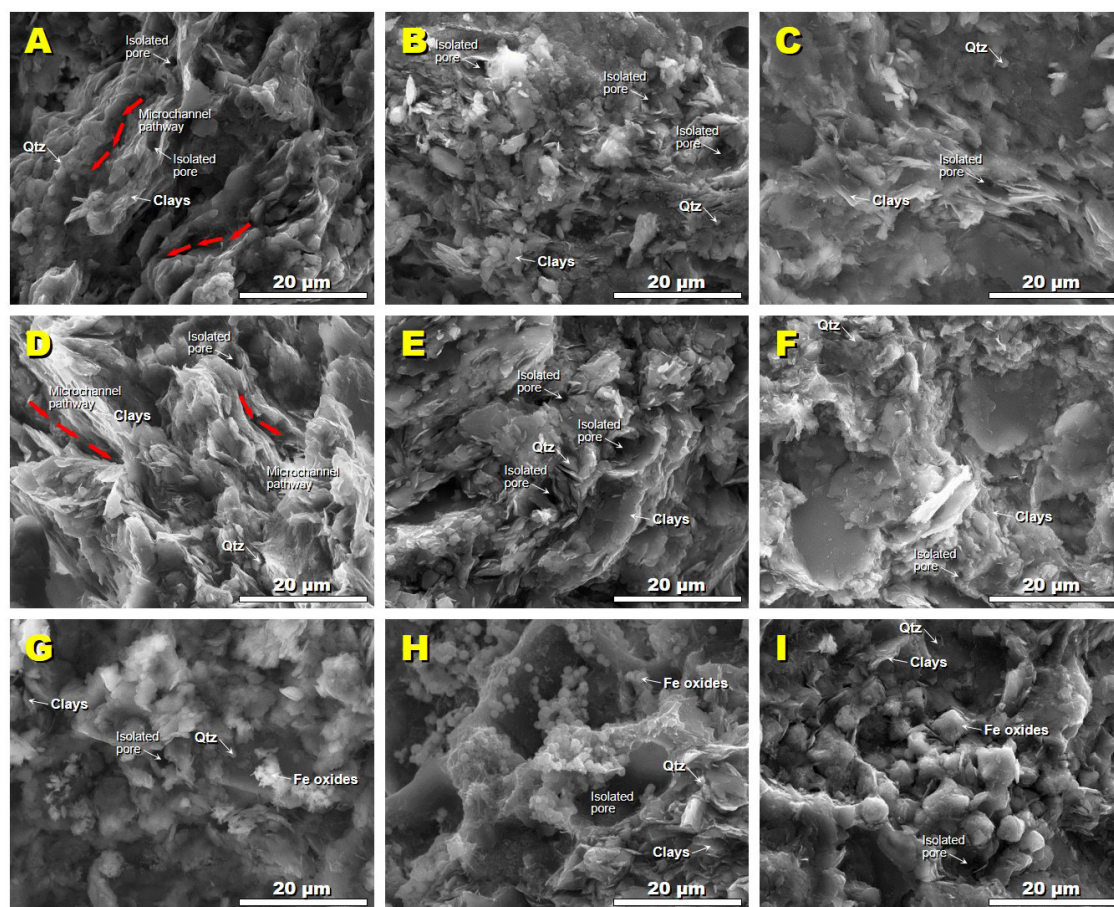


FIG. 10. SEM images of fractured surfaces of the FLO-16 sample in the LSI, Floresta Formation (see Table 1 for the XRPD results). **A-F.** Typical “cardhouse” structure of individual edge-face- or edge-edge-oriented flakes of clay particles, blocky morphology of quartz, and isolated pores; note the occurrence of microchannel pathways in A and D. **G-I.** Isolated pores between clay particles, partially filled by spherical and blocky iron oxide minerals and quartz.

usually bordering quartz grains. Iron oxides (hematite) and oxyhydroxides (goethite) are also present in some parts (Fig. 10G-I), which can be attributed to weathering processes. Quartz grains are ubiquitous and usually display blocky morphologies. The most common accessory mineral phases are anatase, zircon and monazite.

The rock-forming minerals of the siltstones of the Floresta Formation were investigated by semi-quantitative EDS analysis (Fig. 11). We identified ten major constituent minerals, whose EDS signals are explained in detail here. The EDS spectrum for quartz is characterized by the intense peaks of Si and O; where the presence of Al, K and Mg is attributed to adjacent aluminosilicate minerals. The

EDS spectrum for kaolinite is characterized by nearly similar peak heights of Si, Al and O; with K and Fe associated to adjacent illite. The EDS spectrum for illite yields the major elements: Si, Al, and K, with a minor amount of Mg and Fe. Note that the relative peak height of K is less than that of Al, which contrasts with K-feldspars where both K and Al peaks are of equal height. The EDS spectrum for smectite shows the main elements Si, Al, K, Ca, Mg, Fe, and O. Interpretation of EDS spectra of mixed-layer minerals such as illite-smectite is difficult due to electron beam penetration through the clay minerals into the underlying substrate, yielding a composite EDS spectrum (Welton, 2003). The EDS spectrum for chlorite yields the major

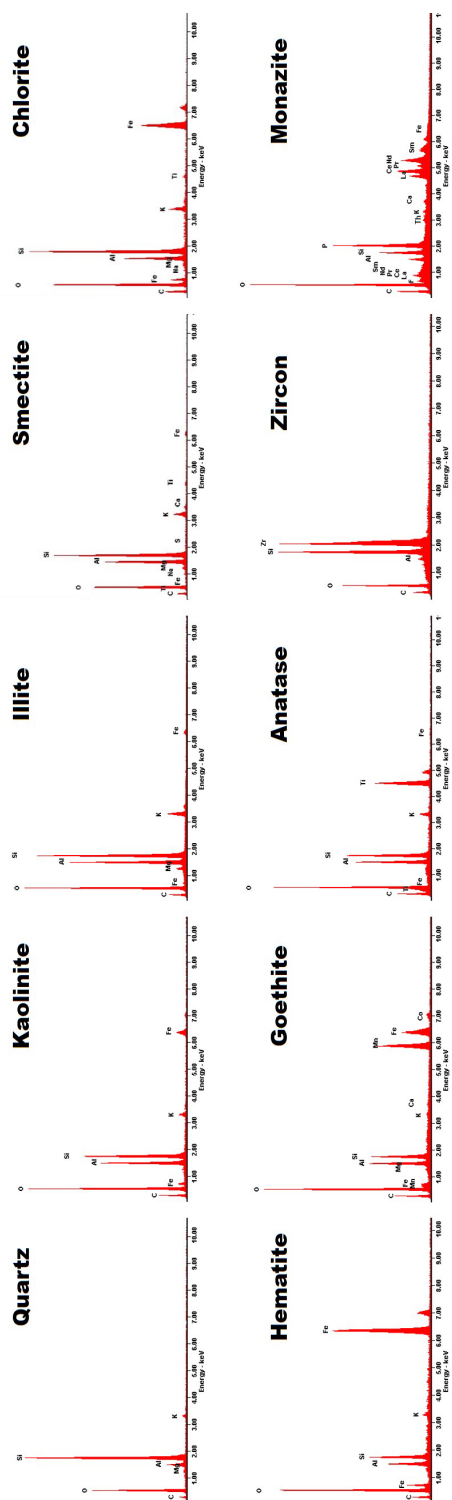


FIG. 11. SEM/EDS spectra of the ten major rock forming minerals of the FLO-13 sample, Floresta Formation (see text for details).

elements Si, Al, Mg, Fe and O; the amount of Fe being highly variable although in this analysis the sample is Fe-rich. Hematite can be recognized in the EDS spectrum by its characteristic Fe and O peaks, whereas Si, Al and K can be attributed to adjacent aluminosilicate minerals. The EDS spectrum for goethite exhibits the characteristic peaks of Mn, Fe and O; with the presence of Mn attributed to very fine inclusions within goethite and/or accumulated on its active surface (Baioumy *et al.*, 2013). Anatase is characterized by the presence of O and Ti elements in the EDS spectrum; the peaks in Si, Al, K, and Fe are not constituents of this oxide but contaminants from adjacent aluminosilicate minerals. The EDS spectrum for zircon reveals the presence of Si, O, and Zr. Finally, the EDS spectrum for monazite is characterized by the presence of Th and lanthanides, with nonradioactive accompanying elements, such as P and Ca, which are distinct markers for this mineral phase.

5. Discussion

According to sedimentological, stratigraphic, and fossil analyses, the sedimentary rocks of the Floresta Formation were deposited in an epicontinental, transgressive and regressive marine environment at the edge of the Paleo-Tethys Ocean. The age of the Floresta Formation has been estimated to be Late Emsian to Early Givetian (Morzadec *et al.*, 2015). The mineral composition of these rocks shows a predominance of silicate minerals, with only a few calcareous rocks developing thin lenses in the LSI. These rocks are also characterized by a flocculated to laminated microstructure, product of the original deposition environment and subsequent diagenesis. Mechanical compaction produced an anisotropic rearrangement of clay minerals (Table 1; Fig. 10).

5.1. Environmental conditions of the Floresta Formation

The recognized fossil fauna suggests that the Floresta Formation was deposited under shallow-water shelf conditions, with eustatic sea-level fluctuations, which would explain the presence of carbonate lenses interbedded with siliciclastic mudstone layers. We consider that there is no evidence of a single transgressive event as proposed in previous studies (*e.g.*, Pastor *et al.*, 2019), but

the transgression-regression events identified here would culminate in a major regression in the Middle-Late Devonian (*e.g.*, Moreno-Sánchez, 2004; Gómez-Cruz *et al.*, 2015). It has been suggested that the edges of this epicontinental sea were made up of siltstones, claystones and occasionally limestones (*e.g.*, Chicangana and Kammer, 2013). This coincides with field evidence and laboratory analyses that indicate that the rock succession of the Floresta Formation was deposited in a low energy environment on the edges of the epicontinental sea, where a relatively stronger influence of siliciclastic material suggest slight sea level variations.

Although it is not the objective of this work, it can be useful to analyze the paleogeographic position of northwestern South America in the Devonian (Fig. 12; Scotese, 2014). This allows establishing possible paleoenvironmental conditions could explain the formation of sedimentary deposits mainly under marine conditions (Mojica and Villarroel, 1984). It is also important to mention that the Middle Devonian rocks reported to the south and north of the study area in the Quetame (south) and Santander (north)

massifs, are calcareous rocks and mudstones deposited in similar epicontinental marine environments with fluctuations in sea level (*e.g.*, Nance and Linnemann, 2008; Nance *et al.*, 2010; Blandón, 2019; Palacios, 2021). In the Floresta Formation the calcareous levels are scarce and restricted only to lenses within the siliciclastic siltstone's layers.

The Floresta Formation sedimentary sequence in the Floresta Massif can be correlated with the Middle Devonian outcrops of the Santander Massif (Mantilla and García, 2018). For example, to the north of the Municipality of Mogotes, the protoliths of the metasedimentary rocks are siliciclastic sandstones and siltstones, where paleofauna is scarce or not present. Further to the north, Middle Devonian rocks corresponding to the Floresta Formation have been reported near the Municipality of Labateca (Villafañez, 2012). These rocks mainly consist of fine- to medium-grained siliciclastic sandstones with some intercalations of mudstones. These facial changes could be interpreted as a lateral variation from shallow-water shelf to coastal conditions with a higher continental influence in the northern

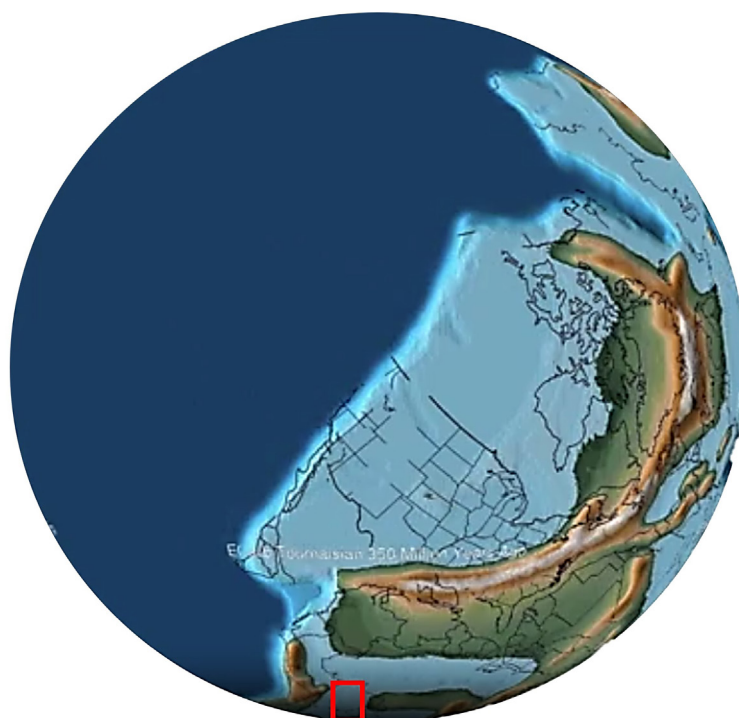


FIG. 12. Middle Devonian paleomap (south polar view). The red square indicates the possible location of northwestern South America. figure modified from Scotese (2014).

sections. Compared to the Floresta Massif, the fossil content identified here could be indicative of warm paleoclimatic conditions (Blandón, 2019), although northwestern South America seems to have been at high southern latitudes ($>50^{\circ}$ S) in the Middle Devonian (Fig. 12), so more evidence is needed to support our interpretations.

5.2. Structural deformation: style and timing

The total thickness of the three intervals measured in the field is ~ 210 m (Fig. 8). The individual thicknesses were checked in three stratigraphic sectors (Teneria and Monticelo, Tablón, and Chuscales and Tobasia; Figs. 5-7). The total thickness is below previous estimates (e.g., Botero, 1950; Mojica and Villarroel, 1984; Ingeominas, 2000; Mozardec *et al.*, 2015). The interpretation of the structural architecture shown in figures 2-4, and the illite (001) XRPD peak shown in figures 9 and 13 suggest that the rocks of the Floresta Formation were affected by burial and intense deformation, accompanied by the development of tight and recumbent folds. This structural history can explain the large difference between the thickness measured in the present study (~ 210 m) and the range of thicknesses (~ 400 - 600 m) measured previously by others (e.g., Botero, 1950; Mojica and Villarroel, 1984; Villarroel, 1984; Barrett, 1988; Morzadec *et al.*, 2015). The repetition of the sedimentary succession by folding, not considered in previous studies, produced an overestimated thickness.

From a tectonic perspective, it is important to note that the underlying Lower Devonian El Tibet Formation does not present such an intense deformation (Figs. 3, 4); neither do the overlying Cucho (Late Devonian), Girón (Jurassic), and Tibasosa (Early Cretaceous) formations (Fig. 3). We suggest that the Floresta Formation underwent a disharmonic deformation (recumbent folds) at a detachment (or shear) level (or levels) that absorbed deformation between brittle layers of the El Tibet, Cucho and Giron formations, possibly due to the rheology of these rocks and the large presence of fluids (connate and/or migrated from other sources). Based on our results, the Floresta Formation seems to have been affected by a complex geological history that includes burial diagenesis, brittle deformation, and associated fluid flow events. These findings are consistent with the geological history of the Floresta Massif (Manosalva-Sánchez *et al.*, 2017).

The structural relationships observed in the field indicate multiple phases of deformation. A first phase of deformation, which postdate the Paleozoic, possibly in the final phase of the Caledonian orogeny, formed the tight folding of the Floresta Formation. A second phase of deformation, sometime during the Mesozoic rifting, allowed deposition along asymmetrical normal fault-related basins, as shown by the lateral changes in thickness of the Floresta Formation identified here and in other localities across the Eastern Cordillera (e.g., Sarmiento, 2001; Kammer and Sánchez, 2006; Sarmiento *et al.*, 2006). A third phase of deformation related to the pre-Andean and Andean deformation, as identified in previous works (e.g., Mora *et al.*, 2010, 2013; Ramírez-Arias *et al.*, 2012). This late deformation caused inversion of preexistent extensional structures and folding of the overlying sequences, and has been identified in the region for Eocene and younger rocks (Ramírez-Arias *et al.*, 2012). The interpretation of low-grade Paleozoic metamorphism in the sedimentary rocks of the Floresta Formation is supported by petrographic and geochronological evidence (e.g., Moreno-Sánchez, 2004; Morzadec *et al.*, 2015; Moreno-Sánchez *et al.*, 2020). The presence of minerals indicative of low-grade metamorphism, such as chlorite and illite (Table 1 and Fig. 9), along with the typical “cardhouse” structure of individual edge-face- or edge-edge-oriented flakes of clay particles (Fig. 10), suggests metamorphic conditions of low temperature and pressure. Based on the IC index shown in figure 13, it is possible to note that a deformation phase occurred near the epizone and upper anchizone boundary (i.e., at a maximum depth of 5-7 km and temperatures around 300°C ; conditions typical of low-grade metamorphism).

Thermochronological studies showed that the Girón Formation, located stratigraphically ~ 1 km above the Floresta Formation, reached a maximum temperature of 165 - 178°C during the Middle Eocene (Ramírez-Arias *et al.*, 2012). Under a normal thermal regime ($30^{\circ}\text{C}/\text{km}$ maximum), the Mesozoic-Cenozoic places the rocks of the Floresta Formation at a maximum temperature of 195 - 208°C . Therefore, the Floresta Formation reached a maximum paleotemperature of $\sim 300^{\circ}\text{C}$ before the Mesozoic. This evidence relates to the first deformation event under regional metamorphism conditions occurred in the Late Carboniferous-Earliest Permian (e.g., Mantilla-Figueroa *et al.*, 2012; Cardona *et al.*, 2016; Van der Lelij *et al.*, 2016).

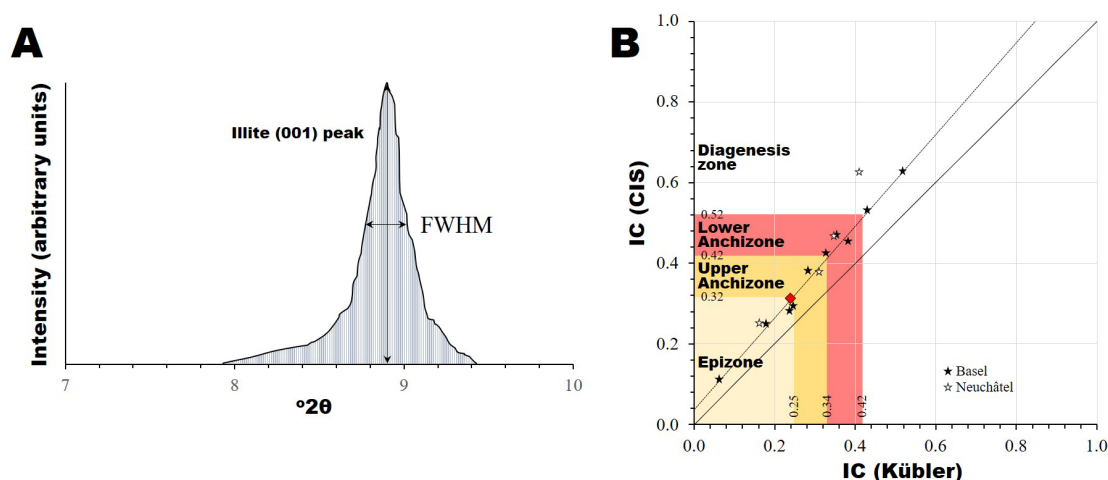


FIG. 13. **A.** Characteristic illite (001) peak of the Floresta Formation rocks (FWHM: full width at half maximum). **B.** IC (Kübler) versus IC (CIS) diagram showing the IC index of the Floresta Formation rocks (red diamond), and the experimental values from the laboratories of Basel (black stars) and Neuchâtel (white stars). Figure adapted and modified after Warr and Ferreiro-Mählman (2015).

5.3. Origin and formation mechanisms of clay minerals

The Floresta Formation mainly consists of siltstones with occasional lenses of limestones (Fig. 8). These rocks are mainly composed of fine granular quartz, illite, kaolinite, smectite, minor chlorite, and I/S interstratified clays (Table 1 and Figs. 9-11). Clay minerals can have different origins and be formed through various mechanisms (*e.g.*, Pittman and Houseknecht, 1992; Meunier, 2005; Robin *et al.*, 2015).

Based on their texture, morphology, and chemistry, we discuss the origin and formation mechanisms of the clay minerals identified in this study. Clay mineral assemblages like illite with minor kaolinite, chlorite, and I/S mixed layers, are often found in sedimentary rock series that have undergone significant thermal influence (*e.g.*, Mangenot *et al.*, 2019). This causes the progressive diagenetic illitization of smectite as temperature increases during burial (*e.g.*, Środoń, 2009). Smectite is highly sensitive to diagenetic alteration (*e.g.*, Dunoyer de Segonzac, 1970; Środoń, 1999). With increasing depth and temperature, smectite is progressively replaced by illite and I/S mixed layers (*e.g.*, Pollastro, 1985; Lanson *et al.*, 2009). The presence of chlorite is common considering that it is a stable mineral phase

at burial conditions, even sometimes associated with very low-grade metamorphism. The minor amounts of kaolinite can be relicts of kaolinite-to-chlorite conversion during the burial process. Kaolinite is unlikely to have been transformed into illite during burial.

According to SEM observations, in the Floresta Formation, clay mineral particles usually exhibit irregular and rounded outlines, indicating a detrital origin, although burial processes could have affected grain morphology as well (Fig. 10). Authigenic clay minerals, like carbonates and chlorites, result from post-depositional diagenesis and possible hydrothermal alteration that supply iron in the system. Detrital clay minerals are weathering products of the initial rock sources and can document paleoclimate variations through the study of their different mineral compositions (*e.g.*, Hong *et al.*, 2017).

Different analytical techniques can yield somewhat different clay compositions. For example, in some cases, the XRPD patterns show the characteristic peaks of illite as the main mineral phase, whereas the SEM/EDS spectrum reveals the occurrence of illite along with kaolinite. This difference could be due to the sample size needed for each technique, implying a non-uniform distribution of minerals within the rock.

The IC index is widely used as a measure of the degree of diagenetic/metamorphic recrystallization of phyllosilicates (Mohammed *et al.*, 2020). Based on the IC index (CIS=0.24; Kübler=0.31) obtained for the Floresta Formation (Fig. 13), it is possible to conclude that they reached upper epizone conditions, near the transition between the epizone and upper anchizone according to the fields proposed by Warr and Ferreiro-Mählman (2015). This indicates a maximum depth of around 5-7 km and a maximum temperature of ~300 °C. The IC values are in the range of those obtained for other rocks elsewhere (*e.g.*, Warr and Rice, 1994; Kübler and Jaboyedoff, 2000; Dellisanti *et al.*, 2008; Warr and Ferreiro-Mählman, 2015), which supports the robustness of our results.

Thermochronology results from adjacent formations, such as the Jurassic Girón Formation, indicate a much lower maximum temperature (165-178 °C), attained during the Middle Eocene. This raises questions about thermal variability in the region and suggests the existence of specific thermal events that affected the Floresta Formation. The extrapolation of the thermal history throughout the Mesozoic-Cenozoic suggests that the Floresta Formation had to experience the significantly high maximum temperatures calculated here before the Mesozoic. This is crucial for establishing temporal connections and correlations with past geological events. The proposed relationship between the maximum paleotemperature of ~300 °C and the first regional deformation event during the Late Carboniferous to Early Permian is intriguing and could hold important implications for the tectonic evolution of the region.

5.4. Fluid-rock interaction and role of clay minerals

The burial diagenesis was followed by alteration due to fluid flow and/or weathering due to exhumation that formed kaolinite and smectite (Figs. 9, 11). Evidence of fluid circulation along rock fractures is inferred by the generation of iron oxide veins, which can be clearly differentiated in the Floresta Formation. However, it is difficult to establish the origin of the fluids and the effects on the composition of the rock. In addition, a carbonate replacement by silicification probably occurred, considering that the edges of the carbonate lenses show replacements by aluminosilicates (Fig. 11).

Iron is contributed to sedimentary environments by weathering processes and volcanic activity. The hydrothermal alteration of the Floresta Formation was tectonically controlled, developing fracture networks (stockworks) with hematite and goethite veinlets, which reflects oxidation probably connected to weathering processes. Hematite is a common accessory mineral in hydrothermal environments, which requires an oxidizing environment to precipitate. It is possible that the amount of iron oxide veins is associated with circulation of fluids resulting from post-diagenetic activity.

The formation of ferruginous veinlets in the rocks of the Floresta Formation is likely related to the activity described by Manosalva-Sánchez and Naranjo-Merchán (2007). According to these authors, acidic volcanic fluids play a crucial role in carrying iron into solution. In the context of the Floresta Formation, the formation of ferruginous veinlets involves the interaction of acidic volcanic fluids with the rock. The acidic nature of the fluids aids in solubilizing iron, which is then transported in solution. Hematite formation may occur as the fluids come into contact with the rocks, causing the iron to oxidize and form hematite-rich veinlets. In a subsequent stage, changes in the ambient conditions, such as pH variations or other environmental factors, could contribute to the formation of goethite. This two-stage process can explain the observed characteristics of the ferruginous veinlets in the Floresta Formation. The partial to complete oxidation of most veins resulting in limonite formation indicates a history of iron-rich fluid interaction.

6. Conclusions

The Early to Middle Devonian Floresta Formation (Floresta Massif, Colombia), has been studied to understand its sedimentology, stratigraphy and paleoenvironmental conditions as well as how it was affected by later tectonic events. The Floresta Formation shows vertical facies changes due to variations from shallow-platform conditions to coastal conditions with a higher continental influence. Unlike the overlying and underlying formations, the Floresta Formation was affected by an intense deformation, accompanied by the development of tight and recumbent folds. This causes the total measured thickness of the Floresta Formation to be ~half as thick as suggested in previous works.

The clay mineral assemblages in the Floresta Formation include mainly illite, with minor kaolinite, chlorite and I/S mixed layers. The illite crystallinity index obtained for the Floresta Formation reveals that it probably reached the upper epizone conditions, in proximity to the transition between the epizone and the upper anchizone, which suggests a maximum depth of ~5-7 km and a temperature of around 300 °C. The burial diagenesis of the Floresta Formation was followed by tectonically controlled hydrothermal alteration, developing fracture networks (stockworks) that were filled with hematite and goethite veins. Iron oxide formation was probably caused by weathering of pre-existing rocks.

Acknowledgements

Authors are very grateful to the Universidad Industrial de Santander for supporting fieldwork at the Floresta Massif (Colombia). The authors are indebted to the laboratories of Microscopy and X-Ray of the Universidad Industrial de Santander at the Guatiguará Technological Park and their responsible professional staff for technical support. We also appreciate the collaboration of G. Bayona for his critical reading and contributions to the manuscript. We are most grateful to the above-named people and institutions for support.

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