Physical and geological description of the Nanchititla dyke swarm

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ABSTRACT

The Late Eocene Nanchititla mafic dyke swarm consists of over one hundred vertical and parallel dyke segments emplaced in southern Mexico. Here, we present a geological and physical description of the dyke swarm, and discuss the emplacement mechanism at the regional and local scale. The host-rock is a continental sedimentary sequence interbedded with volcanic rocks at different levels. Measurements of the apparent Young´s modulus and Uniaxial Compressive Strength (UCS) made with a Schmidt hammer at different levels of the host-rock sequence yielded values from 5 to 45 GPa and from 20 to 90 MPa, respectively. The stratified host rock is mechanically heterogeneous, composed of soft (siltstone) to strong (volcanic breccia) materials. This mechanical contrast has a local effect in the emplacement of dykes inducing sill formation, non-planar morphologies of the dyke walls, en echelon dyke segments, finger-like intrusions, and plastic deformation in some part of the host rock. The weaker siltstone-bearing sequences allowed heterogeneous interaction with the magmatic fluid; whereas stiffer sequences produced fractures parallel to dykes and brecciation. The general trend of the swarm indicates a NNE-SSW orientation of the minimum compressive stress during dyke emplacement, and the parallelism of the dykes with respect to strike-slip faults suggests that the left-lateral deformation regime prevailing in southern Mexico during the early Cenozoic might have influenced its general orientation. We propose that the Nanchititla mafic dyke swarm was emplaced by overpressure of the magma at a shallow crustal level, during the early stages of a non-coaxial transtensional deformation episode in southern Mexico.

Key words: dyke swarm, tectonic regime, apparent Young´s modulus, mechanical stratigraphy, Sierra Madre del Sur, Mexico.

RESUMEN

El enjambre de diques máficos de Nanchititla está conformado por más de un centenar de segmentos de diques verticales y paralelos del Eoceno tardío y se encuentra ubicado en los estados de Michoacán y de México. En este trabajo presentamos una descripción geológica y fisica del enjambre con la cual discutimos el mecanismo de emplazamiento en la escala regional y local. La roca encajonante es una secuencia sedimentaria continental de ambiente fluvio-lacustre intercalada con rocas volcánicas en diferentes niveles. Se realizaron mediciones del modulo de Young aparente y de Resistencia Compresiva Uniaxial (UCS por sus siglas en inglés) con un martillo de Schmidt en diferentes niveles de la roca encajonante, las cuales arrojaron valores desde 5 hasta 45 GPa y de 20 a 90 MPa, respectivamente. La roca encajonante es mecánicamente heterogénea al estar compuesta por materiales suaves (limolita)

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a resistentes (brecha volcánica). Este contraste mecánico influye en el emplazamiento de los diques provocando la formación de sills, morfologías no planares de las paredes de los diques, segmentos de diques en echelon, intrusiones en forma de dedos, y evidencias de deformación plástica en la roca encajonante. Las secuencias débiles de limolita permitieron una interacción heterogénea con el fluido magmático, mientras que las secuencias más rígidas produjeron fracturas paralelas a los diques y brechamiento. La orientación general del enjambre indica una orientación general NNE-SSW del esfuerzo mínimo compresivo, y el paralelismo de los diques con las fallas de rumbo sugiere que el régimen de deformación lateral izquierdo del Terciario en el sur de México pudo haber influido en su orientación general. Proponemos que el enjambre de diques de Nanchititla fue emplazado por sobrepresión del magma en un nivel cortical somero, durante las etapas tempranas de un episodio de deformación transtensional del sur de México.

Palabras clave: enjambre de diques, régimen tectónico, Modulo de Young aparente, estratigrafía mecánica, Sierra Madre del Sur, México.

INTRODUCTION

The Late Eocene Nanchititla basaltic dyke swarm is a regional geological structure constituted by a group of parallel WNW-ESE-oriented mafic dykes (Serrano-Durán, 2005), intruding the upper crust in the north-central part of the Sierra Madre del Sur, southern Mexico (Figure 1). This major structure is relevant in the evolution of southern Mexico, where shearing deformation has been reported in many regions during the early Cenozoic (Menella *et al.*, 2000; Morán-Zenteno *et al.*, 2007; Cerca *et al.*, 2007; Martini *et al.*, 2009). Yet a detailed structural study of this large dyke swarm is lacking. Magmatic and volcanic processes produce fracturing of the crust whose geometry is commonly controlled by the regional stress state, and thus dyke swarms can be used to understand the stress conditions that prevailed at the time of intrusion.

Cenozoic large-scale left-lateral shearing and mid-crustal rock exhumation has been documented by numerous authors along the Pacific coast at the northern boundary of the Xolapa metamorphic complex (Riller *et al.*, 1992; Schaaf *et al.*, 1995; Morán-Zenteno *et al.*, 1996; Tolson-Jones, 1998, 2005; Ducea *et al.*, 2004; Corona-Chávez *et al.*, 2006; Solari *et al.*, 2007; Cerca *et al.*, 2009; Martini *et al.*, 2009; among others). Parallel WNW-ESE shear zones active in Cenozoic times have also been documented north of the Xolapa complex, often associated with silicic magmatism (Alaniz- Álvarez *et al.*, 2002; Morán-Zenteno *et al.*, 2004; Morán-Zenteno *et al.*, 2007; Martini *et al.*, 2009; Martiny *et al.*, 2012). Particularly, several Late Eocene-Early Oligocene silicic volcanic centers, such as Nanchititla, La Goleta, Taxco, Tilzapotla, and Huautla, and left lateral fault systems have been described east of the dyke swarm area (Figure 1) (Alaniz- Álvarez *et al.*, 2002; Serrano-Durán, 2005: Morán-Zenteno *et al.*, 2007; Cerca *et al.*, 2007; González-Cervantes, 2007; Martini *et al.*, 2009; Martiny *et al.*, 2012; Díaz-Bravo and Morán-Zenteno, 2011) (Figure 1). This suggests that magmas were channeled along this weakness zone during a relatively short period of time between the Late Eocene and the Oligocene (Figure 1). The Nanchititla dyke swarm was emplaced at the westernmost end of an alignment of silicic magmatic centers, making the zone an important natural laboratory for the study of these processes. In this paper we present a description of the Nanchititla dyke swarm and address some of the physical conditions of its emplacement.

GEOLOGIC AND TECTONIC CONTEXT

Our study was carried out in the area with the best outcrops of the late Eocene to Early Oligocene Nanchititla mafic dyke swarm (Figure 2a). Other younger dykes, oriented NE-SW, are also present (*e.g.*, the Tuzantla dyke; Serrano-Durán, 2005) but are not considered in this study because they represent a younger volcanic episode. The dykes crop out along a 32 km long area between the N-S oriented Tzitzio anticline at the west and the Late Eocene Nanchititla silicic volcanic center to the east (González-Cervantes, 2007). They form a 24 km wide swarm oriented WNW-ESE at elevations ranging from 350 to 900 m.a.s.l. The Nanchititla dyke swarm yielded $^{40}Ar/^{39}Ar^*$ ages around 37 Ma (Serrano-Durán, 2005) and some of the larger dykes of this swarm feed the lava flow basalts that predate Oligocene silicic magmatism in the zone. Basaltic flows are partially eroded and their remnants crop out at elevations greater than 1200 m.a.s.l. The host rock is constituted by a thick sequence (more than 2500 m) of conglomerates, sandstones, and siltstones deposited in a continental fluvial and lacustrine environment. The dyke swarm is bounded to the west by the N-S trending Tzitzio anticline of latest Cretaceous to Early Paleocene age (Martini *et al.*, 2009) (Figure 2b). The Nanchititla dyke swarm is oblique with respect to this shortening structure that affected the lower part of the host rock sequence (Figure 2c).

Summary of the stratigraphy and geological evolution

The stratigraphy of the studied area has been described by numerous authors with a major emphasis in the Cretaceous sequences. A critical review of the literature has

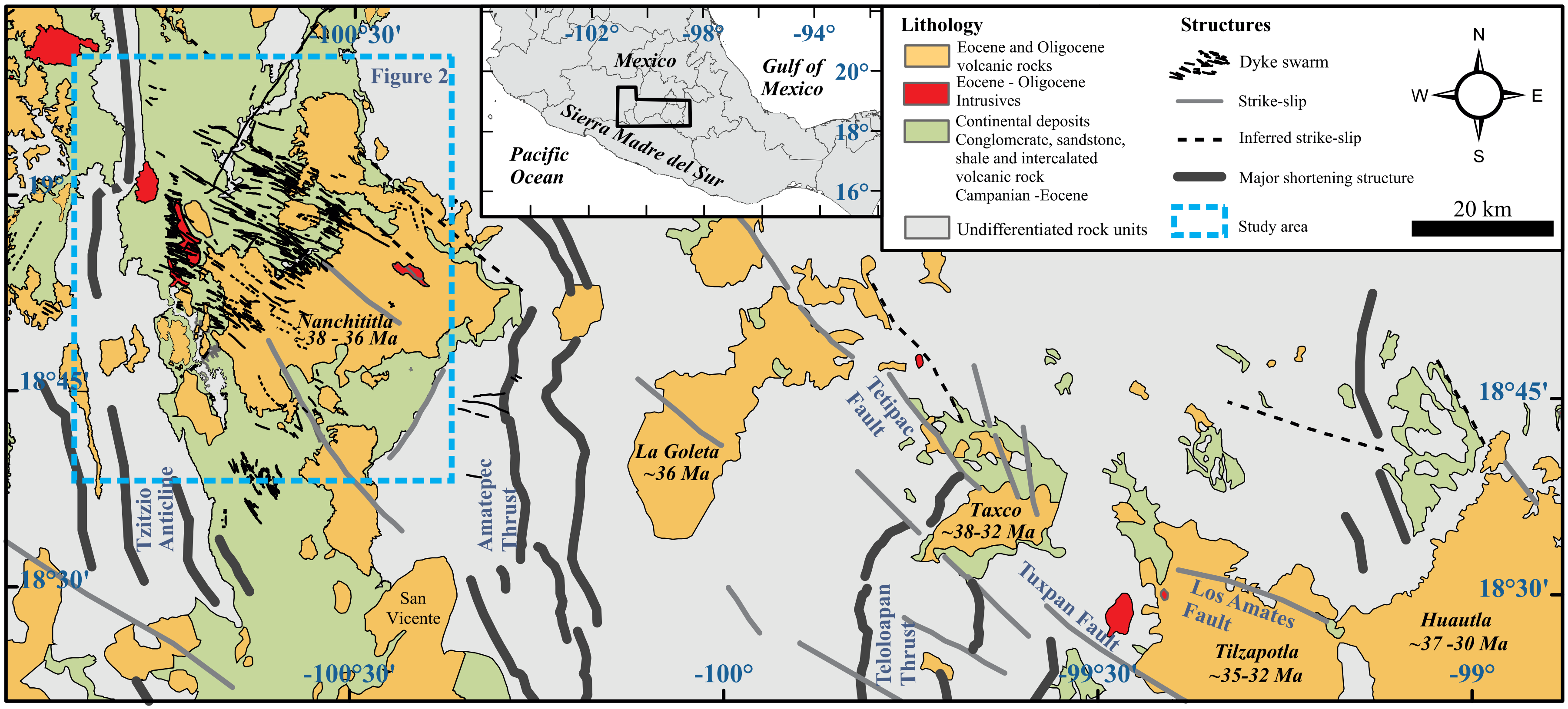

Figure 1. Schematic map showing the regional geological setting. The mafic dyke swarm under study is located at the western end of a regional lineament of Eocene –Oligocene volcanic centers delineating a crustal scale left-lateral shearing zone (modified from Morán-Zenteno *et al.*, 2007). The light blue rectangle marks the location of the study area shown in Figure 2a and the dyke swarm extent.

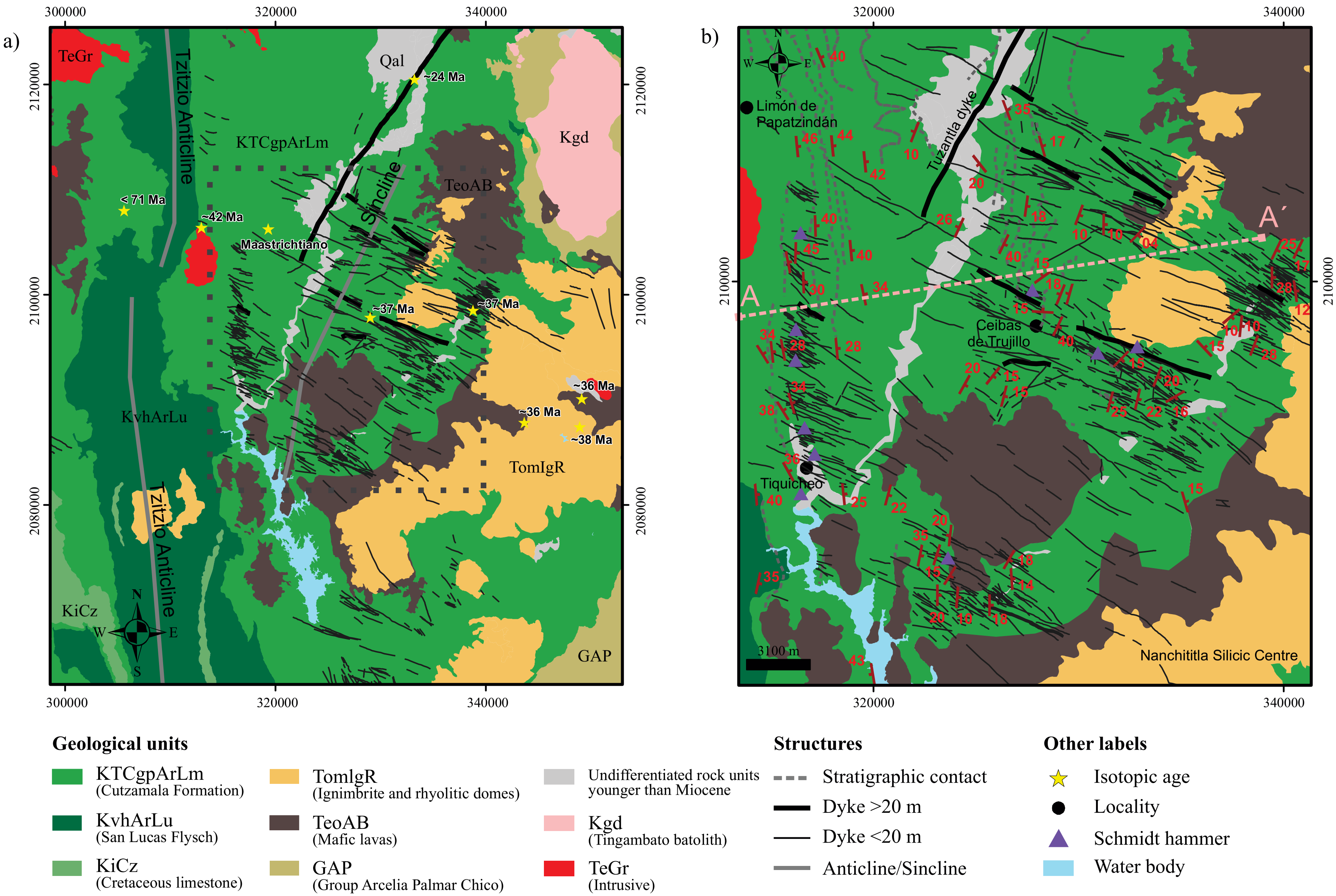

Figure 2. a) Simplified geological map showing the distribution of the dykes (ages and map after Montiel-Escobar *et al.*, 1998, 2000; Serrano-Durán, 2005; González-Cervantes, 2007; Martini *et al.*, 2009; Cerca *et al.*, 2010; and our own mapping); b) close-up of the dyke swarm area enclosed in the dashed square in Figure 2a, showing the distribution of measurements made with the Schmidt hammer (triangles). Geological units younger than Miocene are grouped for simplification; note that some of the eroded dykes, such as the Tuzantla dyke, controlled the deposition of these lithological units.

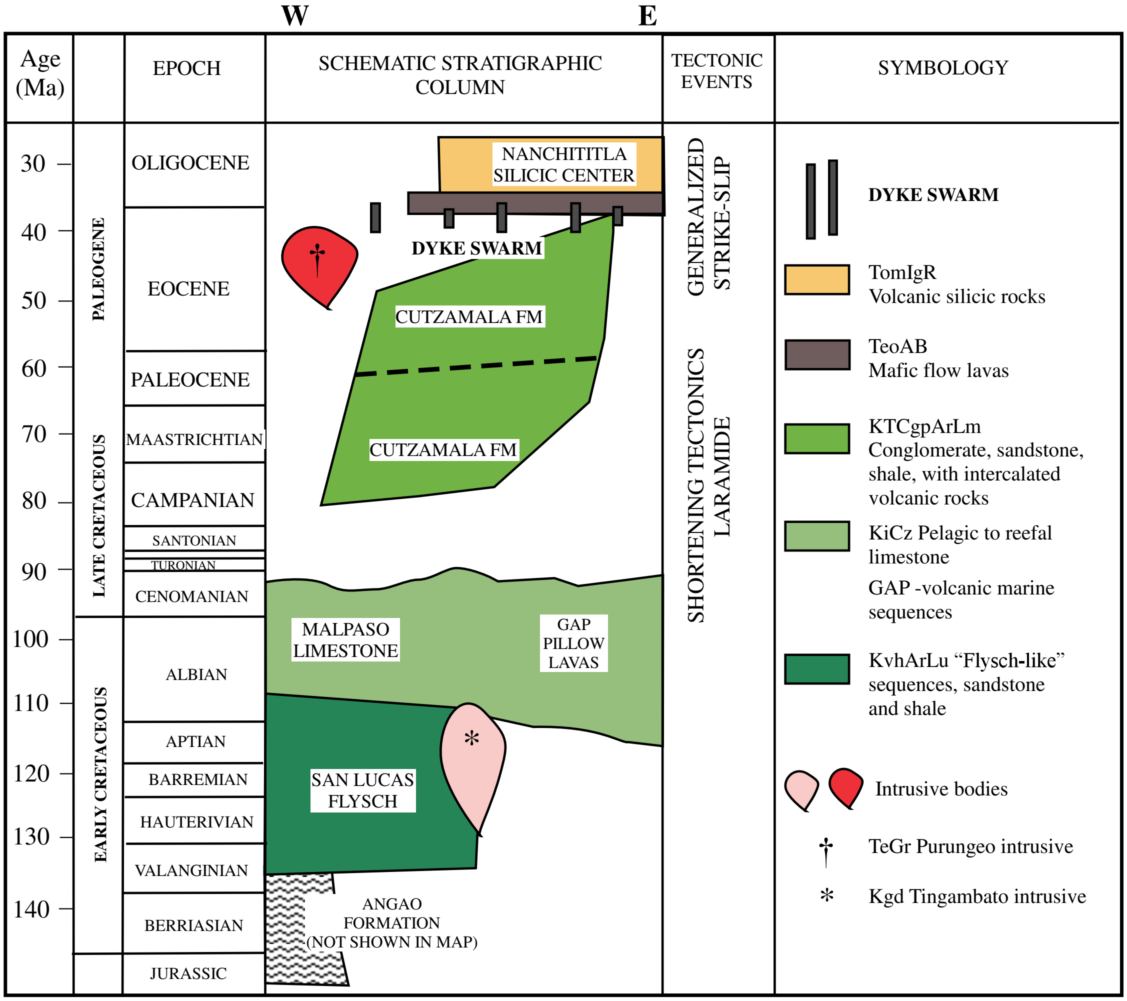

Figure 3. Schematic and simplified stratigraphic column (with data from Montiel-Escobar *et al.*, 1998, 2000; Serrano-Durán, 2005; González-Cervantes, 2007; Mortensen *et al.*, 2008; Martini *et al.*, 2009). The emplacement of the dyke swarm marks the transition from shortening to strike-slip regimes in southern Mexico.

been recently provided by Martini *et al.* (2009) and Cerca *et al.* (2010) and is synthesized here (Figures 2 and 3).

Late Jurassic and Cretaceous succession

The host rock sequence overlies Late Jurassic to Cretaceous sedimentary and volcano-sedimentary rocks of the Arcelia – Palmar Chico group (GAP) in the east and the Huetamo succession in the west (Figure 2) (Guerrero-Suástegui, 1997; Elías-Herrera *et al.*, 2000; Talavera-Mendoza *et al.*, 2007; Garza-González-Vélez, 2007; Martini *et al.*, 2009). In earlier works, these sequences were thought to be part of different crustal blocks or sub-terranes (Campa and Coney, 1983; Talavera-Mendoza and Guerrero-Suastegui, 2000; Talavera-Mendoza *et al.*, 2007), but their lateral continuity below the Late Cretaceous-Early Cenozoic cover has been recently confirmed by Martini *et al.* (2009). In fact Late Jurassic to Early Cretaceous sedimentary and volcanic rocks overlie metasedimentary rocks of Triassic age exposed in the core of the Tzitzio anticline and the Tejupilco anticlinorium, which can be correlated with the Arteaga complex exposed in southeastern Michoacan (Centeno-García *et al.*, 2003, 2008; Talavera-Mendoza *et al.*, 2007; Martini *et al.*, 2009). The GAP consists of massive and pillow lavas intercalated with calcareous sediments (Elías-Herrera *et al.*, 2000), a dioritic-peridotitic body (San Pedro Limón), and serpentinite wedges emplaced along N-S trending faults (Palmar Chico) (Delgado-Argote *et al.*, 1992, 1993). The Huetamo succession is characterized by a gradual transition from pelagic limestone to platform and reefy limestone of Aptian age of the El Cajon Formation (Pantoja-Alor, 1990; Omaña-Pulido and Pantoja-Alor, 1998; Skelton and Pantoja-Alor, 1999; Martini *et al.*, 2009) to subaerial deltaic clastic sediments and biostromic limestone of the Barremian–Early Aptian Comburindio Formation (Alencaster and Pantoja-Alor, 1998; Pantoja-Alor and Gómez-Caballero, 2003). During Albian to early Cenomanian times, carbonate deposition took place at the flanks of the basin represented by the Mal Paso limestone in the Huetamo area (Pantoja-Alor, 1959; Buitrón-Sánchez and Pantoja-Alor, 1998; Pantoja-Alor and

Skelton, 2000; Filkorn, 2002) and the Amatepec limestone in the Arcelia–Palmar Chico area (Elías-Herrera *et al.*, 2000; Cabral-Cano *et al.*, 2000).

Red bed sequence hosting the dyke swarm and the Nanchititla volcanic center

The host rock of the Nanchititla dyke swarm consists of over 2 km of red beds of the Cutzamala Formation (Campa-Uranga and Ramírez, 1979; Guerrero-Suástegui, 1997; Altamira-Areyán, 2002), which covers in unconformity the GAP and the Huetamo succession and, at its base, records an important change from marine to continental deposits. The base of the Cutzamala sequence is exposed in the flanks of the Tzitzio anticline, where a Late Cretaceous age has been established on the basis of paleontological information (Benammi *et al.*, 2005) and geochronological data (Mariscal-Ramos *et al.*, 2005; Martini *et al.*, 2009). The association of sedimentary facies indicates a braided fluvial system discharging through channels in flood plains or lakes (Altamira-Areyán, 2002). Continental sedimentation was partly contemporaneous with regional shortening and continued into the early Paleocene (Martini *et al.*, 2009). The youngest red bed deposits unaffected by shortening are cut by the dyke swarm and other intrusive bodies (Figure 2 and 3) and the uppermost part is partially interbedded and covered by Eocene-Oligocene silicic ignimbrites (Serrano-Durán, 2005, González-Cervantes, 2007; Díaz-Bravo and Morán-Zenteno, 2011). At least two major angular unconformities were recognized in the red beds sequence. We divide informally the Cutzamala Formation in an eastern and western unit, which are separated by an angular unconformity (Figure 2c).

The base of the western unit is composed by massive or crudely bedded conglomerate, gravel, and coarse sand layers. Upwards the unit is characterized by massive, matrix-supported conglomerate and gravel, and interbedded coarse sandstone and siltstone, as well as some paleosoils (Altamira-Areyán, 2002). An andesitic clast from a thick (>2 m) conglomerate has been dated by the $^{40}Ar/^{39}Ar^*$ method at 74 Ma by Martini *et al.* (2009). Occasionally, lava flows and volcanic breccias can be found interbedded with calcareous sandstone and siltstone in the upper part of the sequence. Eastward tilt of this unit varies from 80º to 30º.

The eastern unit is composed mainly by sandstone and limestone layers with interbedded massive conglomerate layers. It rests unconformably over the western unit and the trace of the angular unconformity is roughly parallel to the Tzitizio anticline and approximately marked by the Cutzamala river (Figure 2). Massive coarse sandstone layers less than 1 m thick predominate in this unit. The conglomerate is commonly composed by volcanic, limestone, and sedimentary rock clasts. Interbedded lava flows and volcanic breccias are less common than in the western unit. There are no absolute ages available for the eastern unit, but the fact that the youngest red beds deposit is intruded by the dyke swarm and other intrusive rocks of early Eocene age suggests a Paleocene to Early Eocene age (see Figure 2; Serrano-Durán, 2005).

The dyke swarm is bounded to the east by the Nanchititla volcanic center, which includes mafic lava flows, rhyolitic domes, and thick ignimbrite deposits of late Eocene age (González-Cervantes, 2007). These rocks mostly cover the dyke swarm in the southeastern part of the studied area, although some dykes with WNW-ESE orientations have been also observed intruding the silicic rocks.

Tectonic structure of the region

Structural studies of the dyke swarm region and surroundings are scarce (Jansma and Lang, 1997; Mennella *et al.*, 2000; Serrano-Durán, 2005). Previous to the period of strike-slip shearing, the area underwent shortening during the Late Cretaceous and Early Paleocene that resulted in a pervasive deformation observed at the regional scale in N-S–oriented folding and thrusting (Martini *et al.*, 2009; Cerca *et al.*, 2010). Shortening structures in the area include eastward thrusting and transpression of the Arcelia–Palmar Chico Group (Delgado-Argote *et al.*, 1993) against the Tejupilco anticlinorium and gentle folding in the Huetamo area. The Tzitzio anticline is a major asymmetric fold with an axial plane steeply dipping to the west developed in post-Campanian times (Martini *et al.*, 2009). The sequence of red bed hosting the dyke swarm was deposited within the wide syncline between the Tzitzio anticline and the Arcelia-Palmar Chico group (Figure 2c).

METHODOLOGY

Spatial distribution of the Nanchititla dyke swarm, structural data, and field observations

In this study we complement a structural description of the dyke swarm with an estimation of the elastic properties of the host rock. In a first stage, the available geologic information (Montiel-Escobar *et al.*, 1998; 2000; Mennella *et al.*, 2000; Serrano-Durán, 2005; González-Cervantes, 2007) was compiled at 1:50,000 scale in the INEGI (Instituto Nacional de Estadística y Geografía) topographic maps Tuzantla E14A45, Bejucos E14A55, Tiquicheo E14A54, Limón de Papatzindán E14A44, Palmar Chico E14A65, and Huetamo E14A64. The compiled geologic map was refined during six field campaigns. The surface trace of the dykes was mapped in the field wherever possible, and subsequently completed from satellite images and aerial photographs. Dykes and host rocks were sampled and measured at about 200 sites. Measurements included attitude (dip and dip direction) and thickness. We found some difficulties to measure thickness in dykes wider than 20 meters. Detailed observations of thin sections from 15 dykes allowed the description of fabric and mineralogy.

We also collected stratigraphic data from the host rock, and measured faults, fractures, and folds in the surroundings. The description of the sedimentary rocks collected for density measurements was made from hand samples using a stereoscopic microscope.

Mechanical stratigraphy of the host rock using the Schmidt hammer

The elastic properties of the different stratigraphic horizons intruded by the magma, such as elastic modulus (Young, Poisson, and Stiffness) are crucial information to understand the mechanics of dyke emplacement (Pollard and Mueller, 1976). The apparent Young´s modulus provides a realistic and relative evaluation of the stiffness of the host-rock. Assuming that the rocks behave as a linear elastic solid before brittle failure, the stiffness relationship between the different types of lithology within the host rock should be maintained with increasing confinement. The assumption of lineal elasticity before rupture has been considered suitable for this type of studies (Jaeger and Cook, 1979). Estimates of apparent Young´s modulus and uniaxial compressive strength (UCS) were obtained at 31 sites distributed inside the outcrop area of the dyke swarm (Figure 2b). For this purpose we measured the relative surface hardness of the host rock making a rebounding test with the Schmidt hammer, which is widely used for estimating the uniaxial compressive strength and Young´s modulus of rock materials (Basu and Aydin, 2004). This technique consists in pressing orthogonally against a surface with a spring-loaded piston and measuring the rebound of the piston that is produced by the hardness (penetration resistance) of the rock. The time in which the piston rebound is directly related with the penetration time. The distance traveled by the piston after the rebound (in terms of the initial size of the spring) is called the rebound value R. The obtained value of R is related to the density (void ratio), Young´s modulus, and uniaxial strength of the rock materials. In this work only the rebounds in the horizontal direction were considered in order to avoid gravity effects. Dissipation and loss of energy during the rebound were minimized by selecting the less fractured and most even surfaces. To obtain the elastic parameters of the host rock from the rebound value, we measured in the laboratory the dry bulk density of the host rock measure point on a hand sample collected at each site. The bulk density was determined by two independent methods: 1) the mass of the sample was obtained directly in a digital precision scale and the volume by measuring the amount displaced when introduced in a recipient full of mercury; 2) the mass of the sample is compared in air and in distilled water using a Radwag Specific Gravity Measurement Kit. The density values of the samples were then used to calculate the elastic parameters using the rebound measurements and the empirical relations of Aufmuth (1973) and Aydin and Basu (2005).

RESULTS

Regional observations

Most dykes from the Nanchititla swarm are exposed laterally for 1 to 4 km in length (Figure 4a). Similar to fracture networks, dyke swarms typically display thickness and size distributions that follow a power law (*e.g.*, Gudmundsson 1995a, 1995b; Gudmundsson, 1983; Geshi *et al.*, 2010). Seven major dykes with widths between 20 m and 45 m are found in two WNE-ESE parallel bands (Figure 2b and 4b). Dykes between these two bands are thinner than 20 m (up to less than 1 m) and tend to be oriented mostly asymptotically with respect to the main tendency (Figure 4b). The noticeable exception to the general WNW-ESE trend is the wide (32 m) and long (25 km) Tuzantla mafic dyke, which is almost perpendicular to the main trend, striking NE-SW. However, preliminary age determinations for this dyke (Ferrari and López, unpublished data) yielded an Early Miocene age, which along with its constrasting orientation allowed us to separate it from the Nanchititla dyke swarm.

In the field the dykes are seen intruding only the red beds sequence. At the present depth of exposure none of the observed dykes occupy pre-existing faults or fractures that might have significantly displaced the red beds before

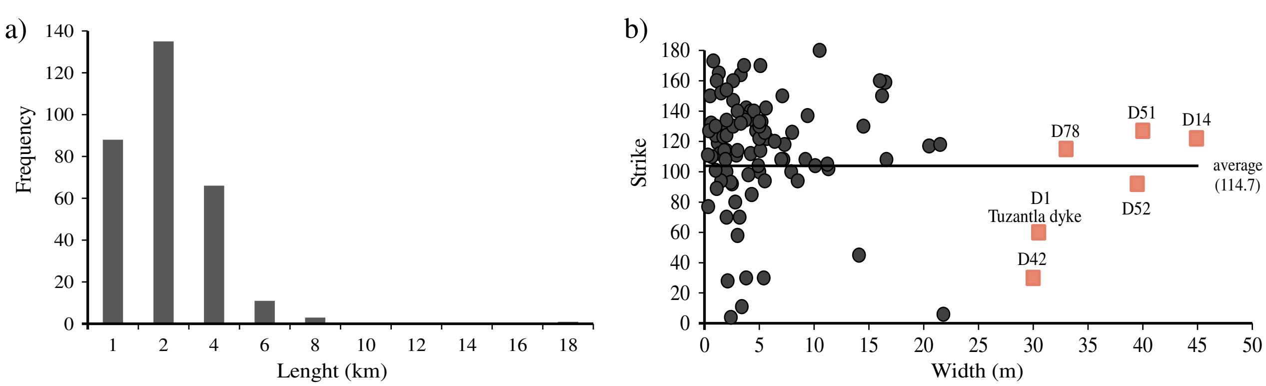

Figure 4. a) Distribution of the apparent length of dykes exposed on the surface; b) relation between strike and width of dykes.

the dyke intrusion. This can be observed by close inspection of the main stratigraphic levels of the host rock where vertical displacements are less than 10 m, as indicated by stratigraphic markers on both sides of dykes (see Figure 2a). On the other hand, the general trend of the dyke swarm coincides with the general WNW-ESE trend of the shear zone of southern Mexico (Figure 1).

Morphology of dykes

Many dyke walls are parallel surfaces that are presumed to have opened symmetrically. However, walls with non-planar morphology were often observed at different stratigraphic levels of the host-rock. In this group of structures we include the case of *en echelon* geometries of dykes that appear dissected or apparently displaced in segments along the stratigraphic joints (Figure 5a, 5b and 5c). At these outcrops we observed that magma linked a pair of *en echelon* dykes with dextral displacement (Figure 5d). In these cases the segments do not overlap vertically. The offset, measured orthogonal to two equivalent walls in each dyke segment are 100 cm, 90.4 cm and 115.2 cm, respectively, and are larger than the width of the segment, for the three non linked segments in a, b and, c in Figure 5. By contrast, in the case portrayed in Figure 5d (from right to left) offset were 177.3 cm and 112 cm, *i.e.* shorter than the dyke width. Commonly, the growth of different *en echelon* segments result in overlap distances much greater than the offsets (Pollard *et al.*, 1982). The growth of the segments

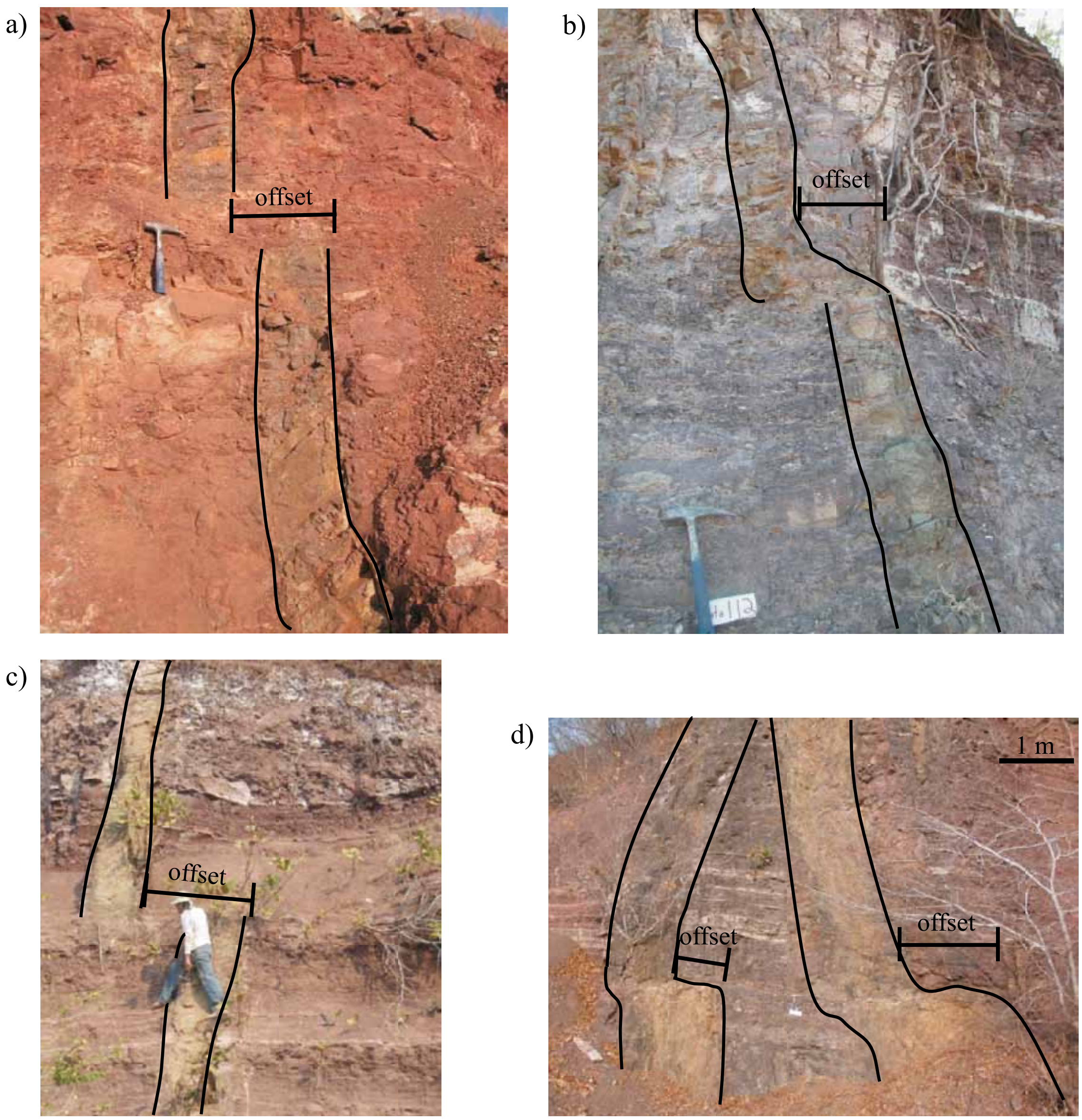

Figure 5. Examples of *en echelon* dykes outcropping vertically in different points of the dyke swarm zone. The continuous line indicates the dyke wall (see text for discussion).

and its subsequent interaction usually produces structures known as horns or bayonets, if a continuous feeding of magma is provided and the compressibility of the intervening host rock does not limit their extension (Smith, 1987). In the present case, the stratigraphic joints acted as surfaces controlling or limiting the formation of *en echelon* segment during dyke propagation.

The walls of eight dykes exhibit a sinuous pattern that increase or decrease locally the width of the dyke (Figure 6). Some of these dykes generated small secondary, finger-like intrusions, with irregular shape and variable width and length (Figure 6b, 6c, 6d, 6e, 6g) that follow an oblique trend but are geometrically concordant with respect to the main intrusion. In one case the finger seems to oppose the direction of the main intrusion (Figure 6c). Only in one case the finger intrusion might be considered as a horn or bayonet that could have been produced by the interaction of two *en echelon* segments (one extending to the surface and the other extending in the opposite direction, Cadman *et al.*, 1990) (Figure 6d). In the rest of the cases, the intrusion pattern and morphology do not show evidence of the interaction of different segments (Figure 6b, 6c, 6e) and, on the contrary, finger intrusions are clearly observed to arise from the main body of magma. Finally, it is common to observe dyke splitting in segments (Figure 6b) expressed by propagating fracture tips, confirming that dykes did not occupy preexisting faults. This bifurcation likely occurred during lateral strain and/or lateral magma flow, for which other evidences will be discussed below.

Dyke-sill linkages

We observed dyke-sill linkage in exposures at different stratigraphic levels close to the contact between the host rock and the mafic lavas at the base of the Nanchititla volcanic center. The walls of the sills are poorly exposed, which hindered the estimation of their width. The observed ending tips of the sills are rounded, and in one case a small extension of the tip produces a finger shape (Figure 7). Other kind of lateral intrusions that arise from dykes are oblique and sub-vertical, relatively short and thin intrusions that follow a sinuous path with a shape similar to fingers (Figure 8). In contrast to sills, these lateral intrusions were emplaced by fracturing a thin bedded host rock (typical thickness of layers is 30 cm). Contacts between the sills and the host rock layers are concordant, and it is not clear if the sills were emplaced between two specific layers or if they intruded a package of several layers. The formation of sills is often explained by the rotation of σ_3 from horizontal to vertical. The arrest of the magma ascent and the lateral growing of an intrusion are commonly attributed to the reach of neutral buoyancy levels (Lister and Kerr, 1991). Recent investigation in buoyant fractures facing low and high density materials found that layer thickness and overpressure inside the dyke influence the initiation of sills (Taisne and Jaupart, 2009) even at neutral buoyant levels. However, the shallow crustal level of the Nanchititla dyke swarm and the difference in density between basaltic magmas and host rock suggest that reaching a neutral buoyancy level may not be a mechanism to explain the presence of lateral intrusions in this area. Furthermore, sills appear at the same level of dykes, where magma mostly intrudes vertically, indicating that σ_3 is still horizontal (Figure 9). Some recent experimental work shows that sills form when a large compressive stress is applied or when the crack has a small initial volume (Menand *et al.*, 2010). The mechanical contrast of the stratigraphic levels and probably the structures formed in the previous shortening phase might explain the localized formation and scarcity of sills. As has been documented by experiments in dyke arrest and sills formation, the latter might form when magma approaches a competent layer (Rivalta *et al.*, 2005) and high local magma overpressure promote lateral growing (Kavanagh *et al.*, 2006; Kavanagh and Sparks, 2011).

Composition, textures, and petrography

The chemical composition of the Nanchititla dykes varies from calc-alkaline basalt to basaltic andesite (Serrano-Durán, 2005). Some of them are highly meteorized and fractured. The majority of the dykes sampled for this study have intergranular, intersertal, and glomerophyric textures that are characterized by two plagioclase size classes (Figure 10). The matrix is made up by fine-grained plagioclase partially altered to sericite and fine subhedral crystals of pyroxene and olivine. It is common to observe crystal twinning in the plagioclase phenocrysts (Figure 10a, 10b, 10c). Opaque minerals, with the crystalline habit typical of magnetite, may have formed during late crystallization stages and tend to fill fractures in phenocrysts. Plagioclase phenocrysts (from 2 to 2.5 cm) are the dominant phase (Figure 10d, 10e, 10f), but occasionally form glomerophyric textures in association with subhedral and anhedral augite, olivine, and opaque minerals. Hornblende is present as an accessory mineral. In most cases phenocrysts represent over 30 % of the rock. Table 1 shows the content of phenocrysts measured in four selected hand samples. The longest axis of phenocrysts commonly ranges from 0.5 cm to 1.5 cm with maximum up to 2.5 cm. The phenocrysts phase probably acted as suspended particles during flow, as evidenced by many plagioclase phenocrysts that appear broken by collision (Figure 10h and 10i) during high strain rate flow, possibly at the early stages of emplacement. Viscosity of basaltic magmas increases with high phenocryst concentration and can become highly strain rate dependent. In the interval between 20 and 60% of crystal content, flow curves of magmas have a nonlinear behavior (Shaw, 1969; Ryerson *et al.*, 1988). Cracking of phenocrysts in crystal-bearing and low viscosity magmas observed recently in laboratory and numerical experiments (Caricchi *et al.*, 2007; Deubelbeiss *et al.*, 2011) have been associated to high stress conditions during emplacement.

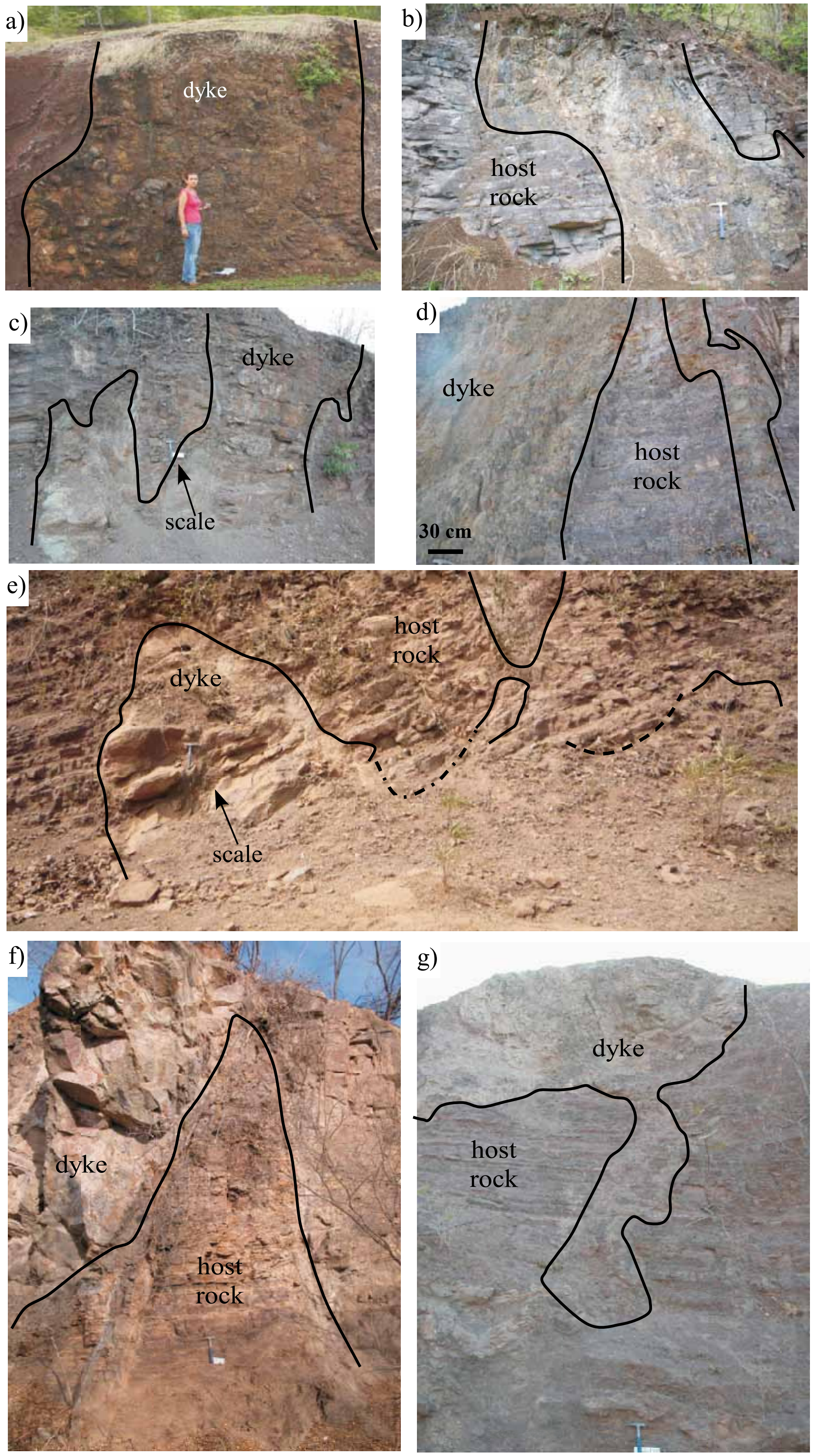

Figure 6. Examples of dykes with sinuous walls. Dykes in b), c), d), e) and g) developed secondary intrusions of finger type. In some of these cases, the main intrusion splits in segments that can follow the finger-like shape as in b), c), d), e), and f).

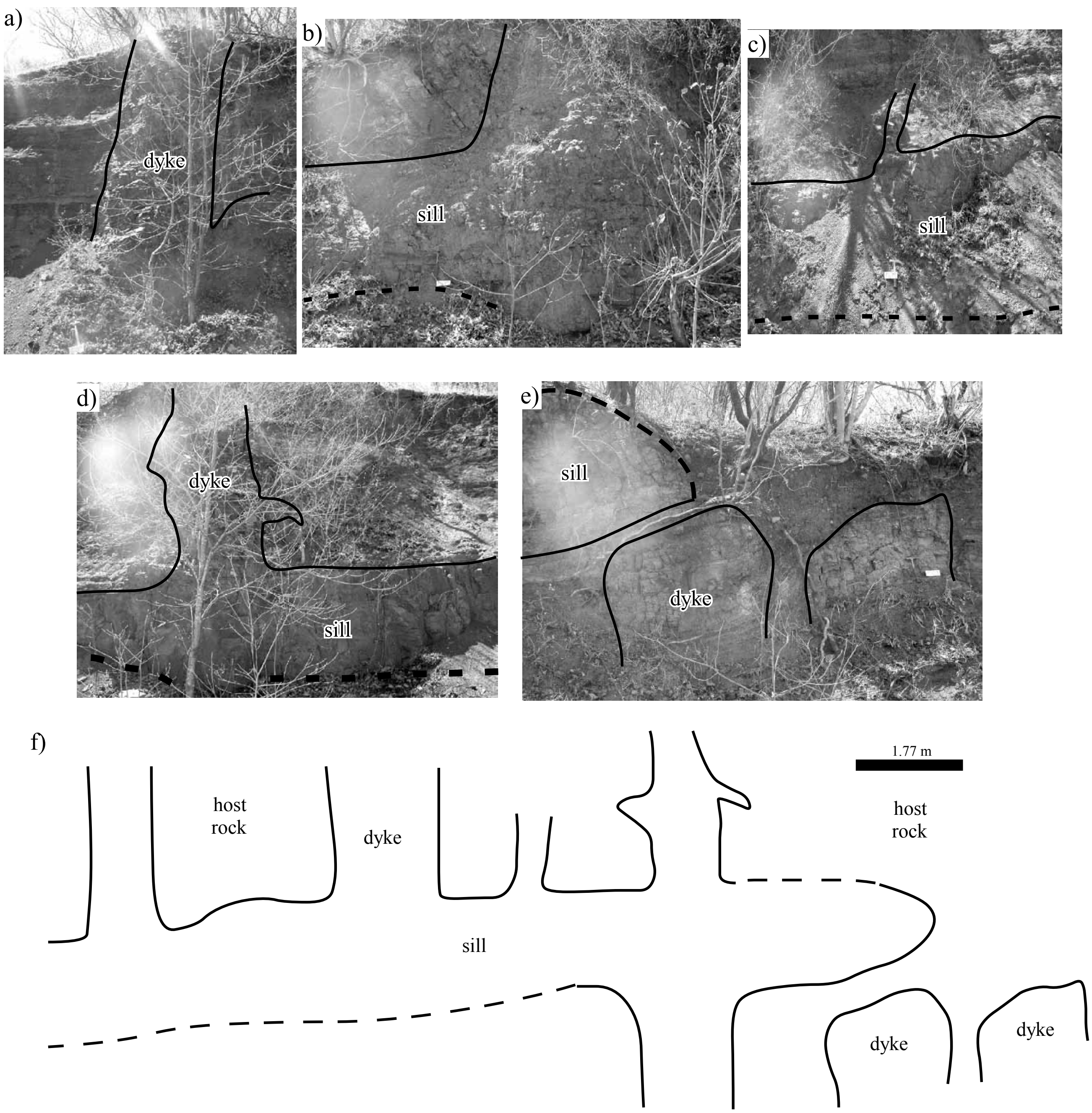

Figure 7. Photographs showing a dyke-sill linkage viewed from left to right. The complete visualization of the system is shown in a scheme in f).

Density and rebound measurements of the host rocks

The *in situ* rebound and laboratory density measurements were obtained for different rock types, which were classified as siltstone, sandstone, conglomerate, and volcanic breccia. Density of theses rocks ranges in the common values for sedimentary rocks from 2.28 to 2.73 g/cm^3 (Figure 11a and 11b). Siltstones show the largest dispersions, ranging from the lowest to the highest values observed for this sequence. Sandstone has a narrow range from 2.4g/cm^3 to 2.6 g/cm^3. In contrast, volcanic breccia and conglomerate have densities above 2.5 g/cm^3.

The Uniaxial Compressive Strength (UCS) is controlled by several inherent and external parameters that are known to affect the mechanical properties such as grain size, packing density, degree of grain interlocking (which refers to the presence of cement, matrix, or welding of the rock), porosity, and mineral composition (Zorlu *et al.*, 2008). For example, rocks containing quartz as a binding material are the strongest followed by calcite, and ferrous minerals; rocks with clayey binding material are among the softest (Zorlu *et al.*, 2008). Since the depositional environment, packing

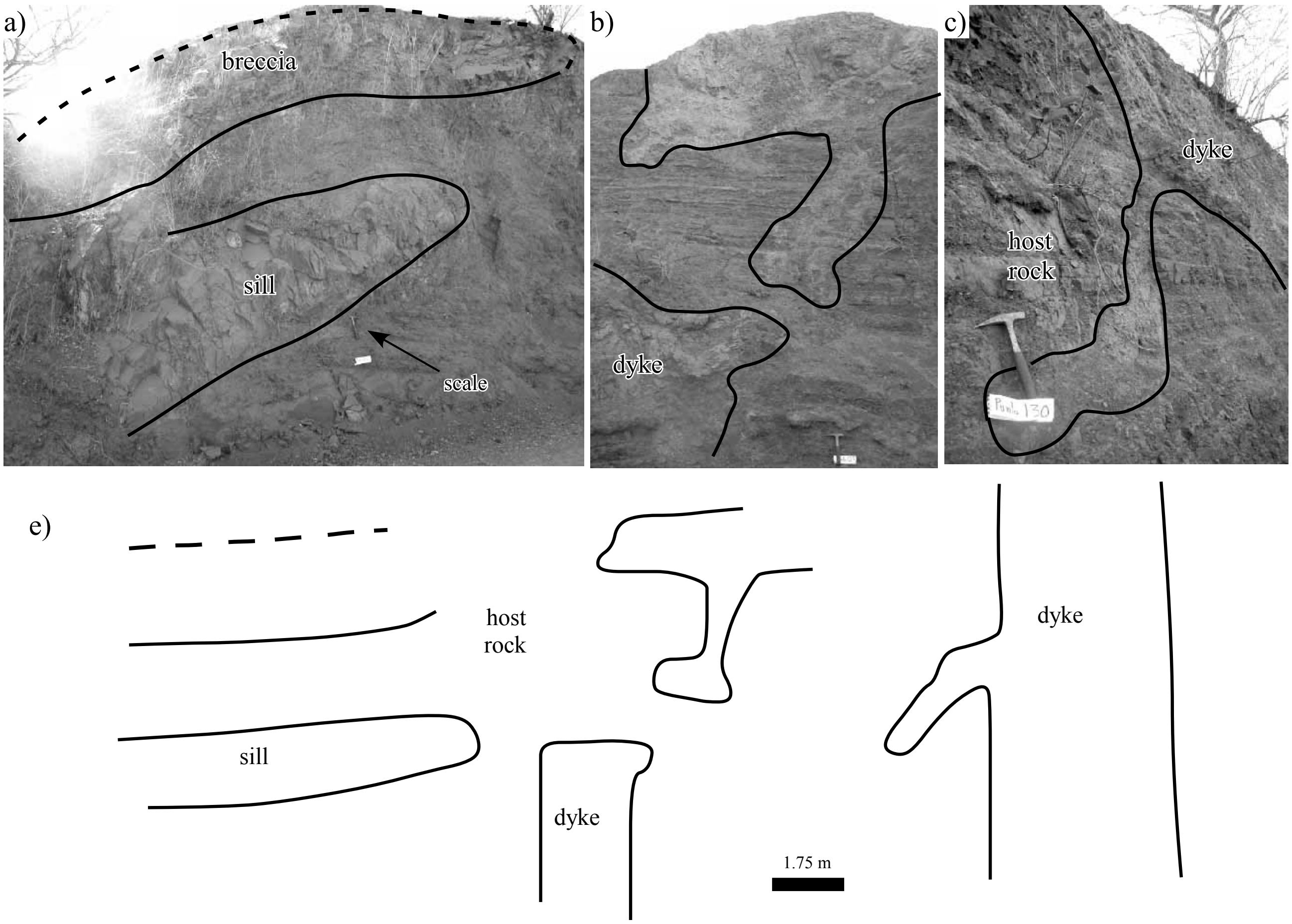

Figure 8. Photographs showing different parts of one example of dyke-sill linkage that follows a sinuous path with a shape similar to fingers. As in the previous figure, a schema of the outcrop is shown in e).

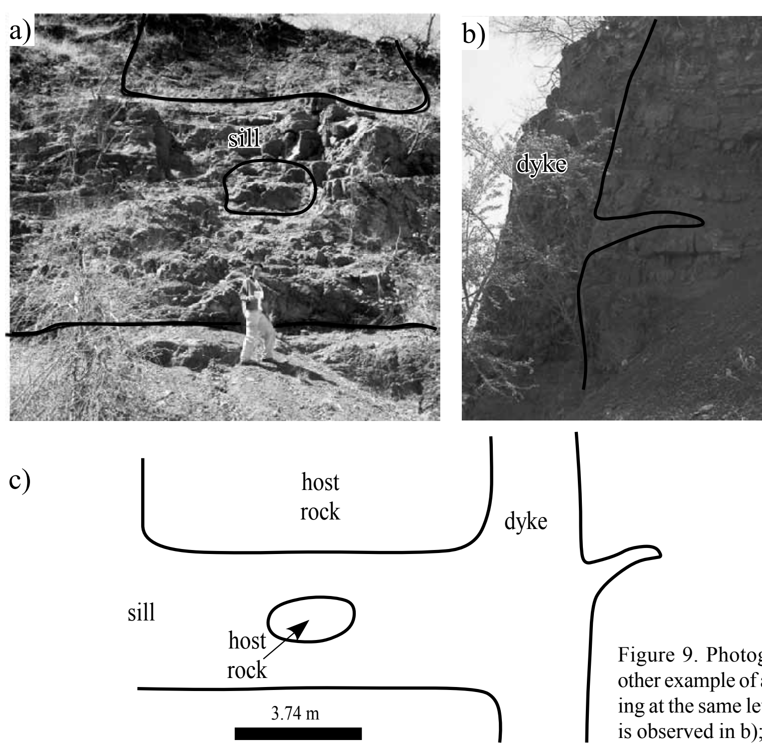

Figure 9. Photographs showing different parts of other example of a dyke-sill linkage with sills appearing at the same level of dykes. A finger-like intrusion is observed in b); an outcrop schema is shown in c).

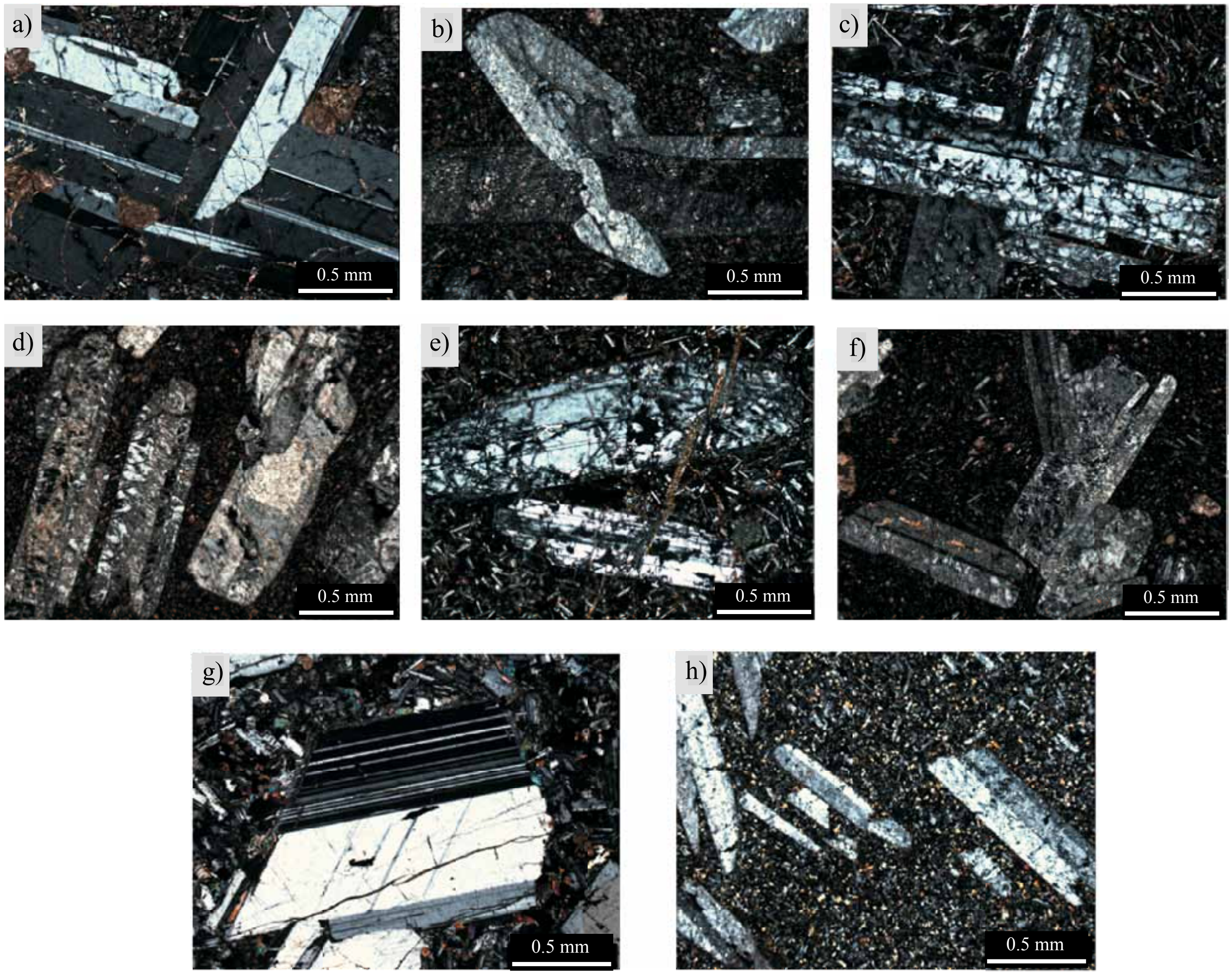

Figure 10. Photographs of thin sections of samples from the dyke swarm. a-c: Examples of twinned plagioclase phenocryst; d-f: examples of fabrics formed by phenocrysts; g-h: fracturing of plagioclase by mechanical interactions.

density, and mineral composition of the sequence are similar in the studied hosting rocks, we expect the mechanical response to be a function of grain size and the degrees of grain interlocking.

The UCS of welded volcanic breccia and cemented conglomerate displays a bimodal behavior (Figure 11c), with a low range of 20 to 50 MPa, and a high peak up to 90 MPa. Clearly the highest value corresponds to volcanic rocks. Siltstone has UCS values below 25 MPa, while sandstone yielded a wider range from very low values of 15 MPa up to almost 60 MPa (Figure 11c). These results reflect the consolidation and, consequently, the high competence of the rock to fracture. The highest values of UCS obtained for the volcanic breccia exceed those of conglomerate (23.57 MPa). On the other hand, siltstone presents the lowest values of UCS while sandstone has intermediate values (Figure 11c). The good differentiation of rocks type and stiffness response is also evident in the Young´s modulus (in GPa), where breccia keeps a wide range from 10 to 45 GPa. Sandstone appears at moderate values between 5 and 32 GPa (Figure 11d) and siltstone keeps below 12 GPa. The results agree with previous estimates of Young´s modulus that range between 5 and 100 MPa for different rock types (*e.g.*, 45 GPa for quartzite, 1–20 GPa for sandstone, 1–5 GPa for shale; Birch, 1966; Pollard and Muller, 1976; Pollard and Fletcher, 2005).

Drag folds, fracturing, and brecciation of the host rock

Drag folds related to dyke walls are often observed in the Nanchititla dyke swarm (Figure 12). Drag folds were observed close (less than 2 m) to the dyke walls in an otherwise horizontally or gently dipping sequence of host rock. We consider that they represent the bending of host-rock before rupture or during over-pressured emplacement of viscous magma that can plastically deform the host rock. During fracturing, materials can undergo some plastic deformation in the region of high stress concentration; for instance near the tip of fractures a plastic weakening zone

Table 1. Modal counting of the phenocryst percentage in four selected samples of the Nanchititla mafic dyke swarm.

Dyke sample	Φ % Face A	n	$A_{max\ phenocryst}$ (mm^2)	Φ % Face B	n	$A_{max\ phenocryst}$ (mm^2)
D8	33	2057	16.99	34	1352	13.73
D48	17	384	127.89	12	66	42.11
D50	17	335	52.90	11	193	71.71
M3D51	38	115	317.5	47	68	172

Φ: Volumetric percentage of phenocrysts measured in a rock plane (face) of about 64 cm^2. n: number of phenocrysts counted. $A_{max\ phenocryst}$ (mm^2): Maximum area of phenocrysts recorded on plane.

has been documented in materials including high strength alloys (Hutchinson, 1968) and dyke host rocks (Sleep, 1998). This deformation is irreversible and is thus recorded in the material after fracturing; it also might play a crucial role in the dyking process that is not yet fully understood (*e.g.*, Correa-Gomes *et al.*, 2001). We observed that drag folds dominantly formed in siltstone and sandstone sequences.

In some outcrops we observed sets of fractures close and parallel to the dyke walls that might have been created during magma emplacement. Figure 13 shows a typical occurrence of fractures associated with dykes. These fractures sometimes are filled with hydrothermal alteration minerals that are commonly associated with the processes of cooling in the dykes (*e.g.* Keating *et al.*, 2008). This type of adjacent fractures has been documented also in kimberlitic dykes (Kavanagh and Sparks, 2011). In a seismic study, shallow intrusions in Iceland have been associated to the deformation produced by a propagating dyke (White *et al.*, 2011). Kavanagh and Sparks (2011) attributed the formation of these fractures to exsolution of gases traveling in front of the tip and to extreme alteration of the host rock, but it is important to note that Late Eocene to Oligocene hydrothermal mineralization is reported in the study region (Camprubí, 2003, Camprubí *et al.*, 2003). Some of these fractures may have remained open in these shallow crustal levels, a fact that reduces drastically the stiffness of the host-rock. No folding, brecciation, or parallel and closely spaced fracturing was observed in the host-rock away from the dykes influence zone.

DISCUSSION

Role of the regional tectonic context in dyke emplacement

The relationship between stress field and the distribution of magmatism at surface has been well documented, and has been used to explain the distribution of volcanoes and dykes swarms (Nakamura, 1977; Gudmundson and Brenner, 2004). Dykes are usually assumed to propagate as magma-driven tension fractures where minimum principal compressive stress (σ_3) is perpendicular to the plane of the dyke (*e.g.*, Pollard, 1987). In extensional environments such as rifts or spreading zones, *e.g.* the Iceland dyke swarms and the Oman ophiolite (Gudmundson and Brenner, 2004), dykes align parallel to the extension axes and generate volcanic cones along their trace. In a regional context, the Nanchititla dyke swarm represents the westernmost known magmatic occurrence associated to the WNW-ESE alignment of silicic centers of southern Mexico; the dyke emplacement started slightly before the emplacement of the Nanchititla silicic volcanic center (González-Cervantes, 2007), and is coeval with voluminous silicic magmatism clearly emplaced along left-lateral strike-slip faulting observed to the east (Alaniz-Alvarez *et al.*, 2002; Morán-Zenteno *et al.*, 2004; 2007; Díaz-Bravo and Morán-Zenteno, 2011; and references therein). Indeed, the Late Eocene to Early Oligocene silicic centers of Taxco (Alaniz-Álvarez *et al.*, 2002), Tilzapotla (Morán-Zenteno *et al.*, 2004), and La Goleta (Díaz-Bravo and Morán-Zenteno, 2011) are probably associated to major bends or step-overs that localize extension and magmatism along the strike-slip regime (Figure 1). These works provide supporting evidence to constrain the overall temporal coincidence between the dyke swarm and the strike-slip regime. The mapped traces of the dyke swarm have an anastomosing pattern with bends and offsets, as well as small dykes diverging locally from major dykes whose kinematic is also consistent with a left-lateral strike-slip deformation. The orientation of small segments of dykes is scattered around the main orientation to which large dykes are parallel, but in general terms the whole dyke swarm follows a WNW-ESE trend. This predominant orientation suggests that magma injection locally modified the orientation of the minimum principal compressive stress (maximum extension) toward an approximate NNE -SSW orientation, producing opening fractures. This observation also implies a partition of the strike-slip strain in a component of shear strain accompanied by dilatation normal to the shear band and the formation of tension fractures. In this context, the lithospheric scale fractures formed during the deformation would open, facilitating the ascent of magma from deep sources to shallow depths. The absence of large normal faults in the region of the dyke swarm also suggests that the volume of magma provided was large enough to accommodate any extension. It is likely that dilatation of the host rock was enhanced by magma overpressure (and possibly vent widening processes such as brecciation) but the overall resulting structural pattern is controlled by the long term and regional shear strain. In summary, the broad

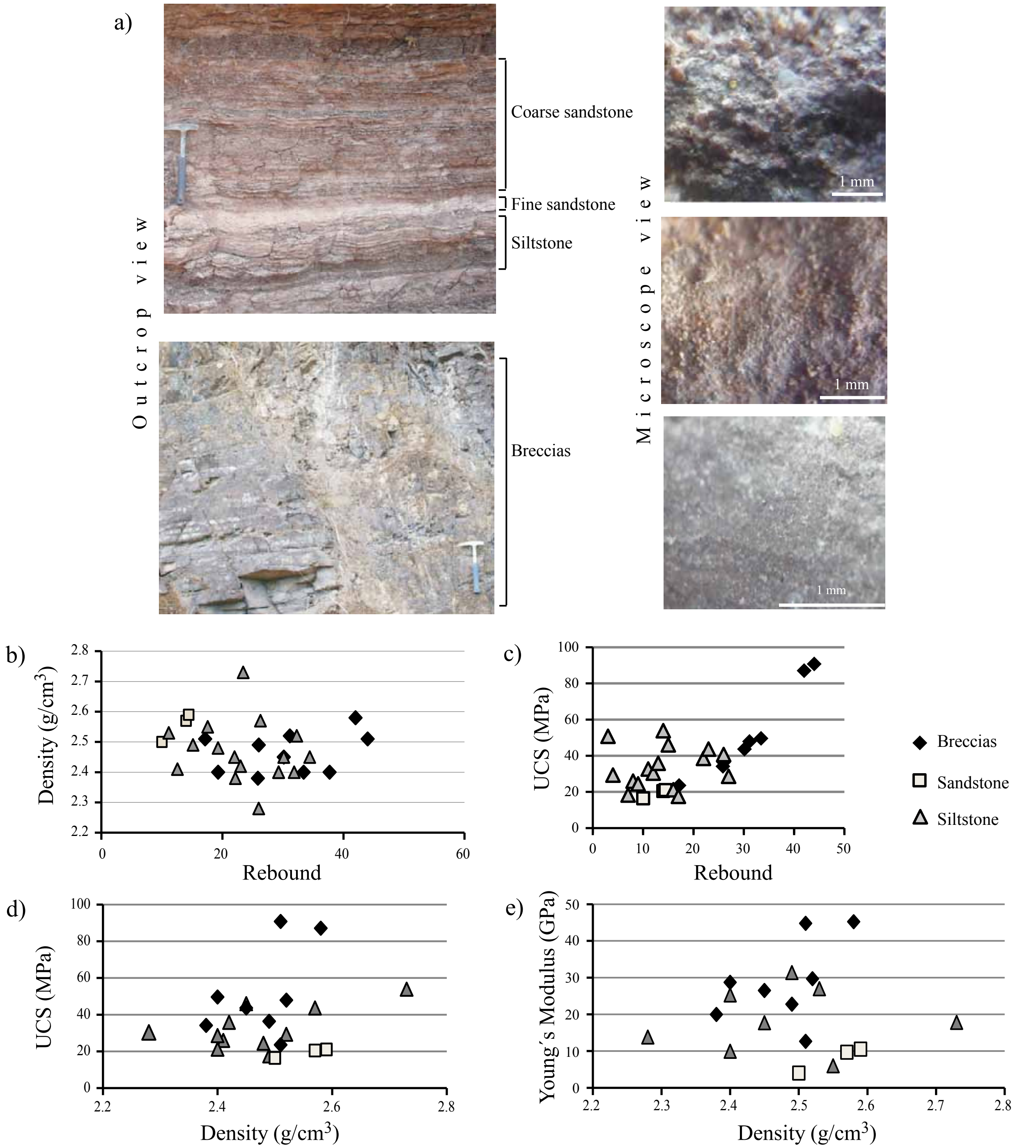

Figure 11. Relative stiffness values of sandstone, siltstone, conglomerate and volcanic breccia in the host rock intruded by the dykes. a: Rebound of Schmidt hammer *vs.* density of the host-rock. b: Uniaxial compressive stress (UCS). c, d: Values of uniaxial compressive stress and Young's modulus in terms of their density; the empirical model of Aufmuth (1973) was used for this calculation.

implication of our model is that the mafic Nanchititla swarm of dykes might have started to emplace during the early stages of the non-coaxial Cenozoic transtensional deformation, and at the western tip of a forming major shearing zone. The Nanchititla dyke swarm share geometric characteristics with other dyke swarms formed under an extensional regime in a previously undeformed succession as the Nandurbar–Dhule dykes in India (Ray *et al.*, 2007), suggesting that an extensional component is necessary to allow the ascent of magma and the formation of dykes. Probably, the initiation of the underlying shear zone controlled the parallelism of the major dykes of the swarm (Figure 14), and the orientation and distribution of major dykes, in turn, determine to a large extent the subsequent dyke pattern. Scholz (2002)

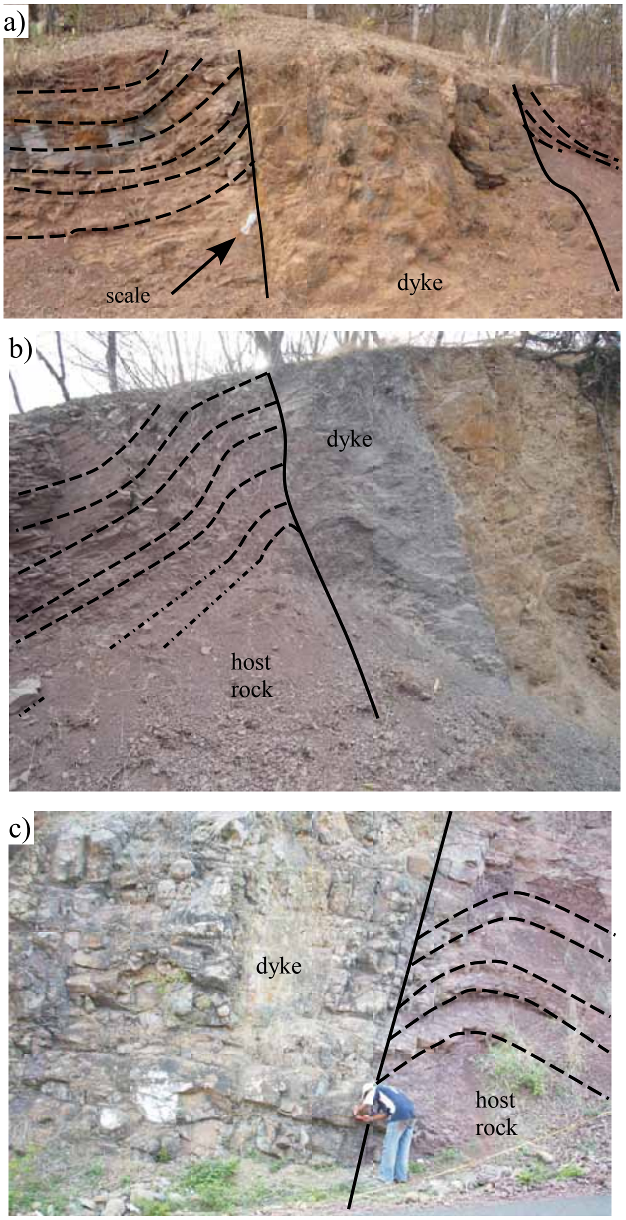

Figure 12. Examples of drag folds close to the dyke walls. The thick and continuous line indicates the dyke wall, dashed lines indicate folded layers.

noted that widespread jointing could be a pre-requisite for the formation of large shear zones. During the time of emplacement of the dyke swarm (37–30 Ma), the central part of the Sierra Madre del Sur offered a scenario where differences in magmatic systems are also evidencing changes in the geodynamic conditions of the crust. We can only speculate on the nature of such conditions that include a number of factors such as the nature and thickness of the crust, different magma chambers or mantle sources, times of residence of the magma in the crust, among others. The Nanchititla dyke swarm represents a unique opportunity to understand the processes that leaded to the present day geological configuration of southern Mexico.

Emplacement mechanism of dykes and local mechanical heterogeneities

The general trace geometry and mode of emplacement of the Nanchititla dykes has been explained in terms of regional tectonic stresses and deformational regime. Wrenching of the host-rock may have resulted in the development of a directional localization of strain softening that permitted the emplacement of overpressured magma causing hydraulic fracturing. Individual dykes are distributed over a wide area (24 km) but we observed the clustering of dykes in at least two narrow bands around the thickest intrusions (dyke 52 and 22). According to Paquet *et al.* (2007), dyke clustering reflects strain partitioning at different levels of the crust that can potentially reflect the location of the magma sources, besides other conditions such as the heterogeneities of the crust and the mechanical response of the host rock. The strength variations depending on the lithology can have an important influence in the style and intensity of deformation (Gross, 1993), resulting from the forces applied during dyke injection. Thus, physical parameters such as the mechanical anisotropy of the host-rock can locally play an important role at smaller scales producing irregular dyke shapes and different emplacement mechanisms that differ from typical sub-vertical planar sheet intrusions.

We have documented emplacement features deviating from the planar intrusion such as finger type intrusions (*e.g.*, Schofield *et al.*, 2010), irregular shape of dyke walls and probably widening of the vent (*e.g.*, Valentine and Krogh, 2006), local change of direction of the emplacement (from vertical to lateral), and dyke-sill linkage that illustrate the importance of local effects in the interaction between dyke and host-rock. We have also quantified the elastic behavior of the host rock related to local emplacement style. The host rock is composed by a group of sedimentary beds and volcanic layers that resulted in a highly mechanically heterogeneous sequence as evidenced by the relative values of UCS and Young´s modulus measured by engineering methods. The relatively high values of bulk density in the host rock indicate that the compressive strength of the rocks resides in grain size and fabric packing. In general, we observed a direct relation between grain size and compressive strength (Figure 13). The highest values correspond to the volcanic breccia that predominates in the western part of the study zone. The lowest values correspond to siltstone located at different stratigraphic levels. From these observations it can be argued that the mechanically weaker siltstone-bearing sequences allow heterogeneous interaction with the magmatic fluid (finger and sinuous dyke walls). On the contrary the volcanic breccia and conglomerate behave in a stiffer manner producing fractures parallel to dykes and brecciation. Further field and experimental examination is needed to detail the complex interactions between magmatic fluid and host rock in this dyke swarm and the influence of other factors not considered in this study, such as thermal effects during emplacement.

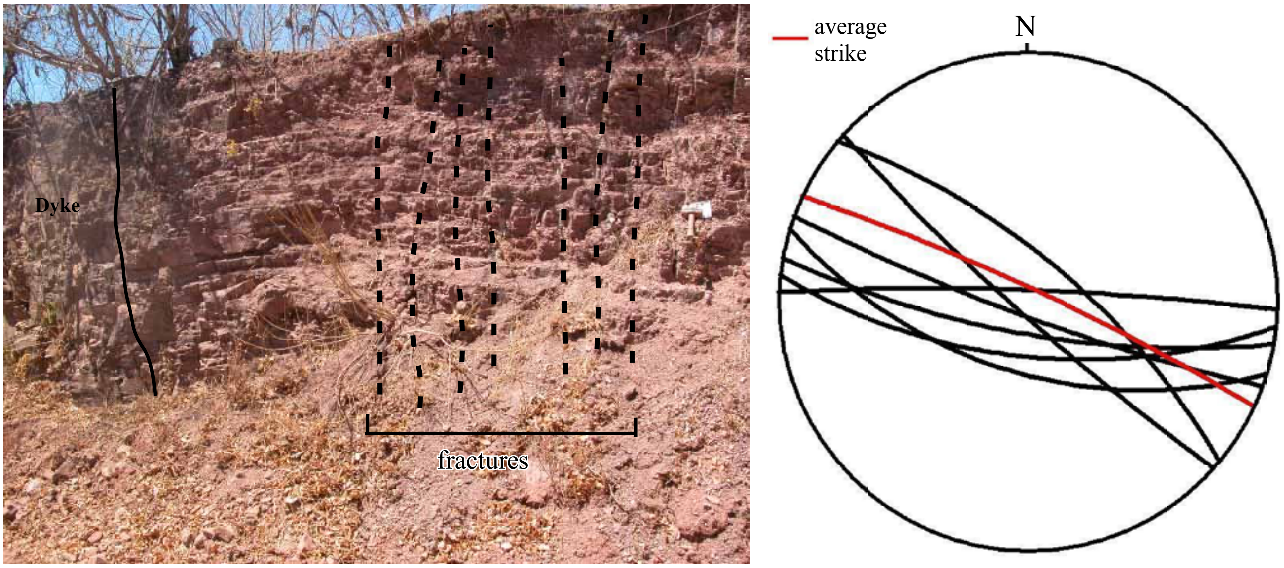

Figure 13. a: Photograph showing an example of a dyke and adjacent fracturing in the host rock. b: Stereogram showing the trend of fractures associated to dykes; the red solid line is the average trend.

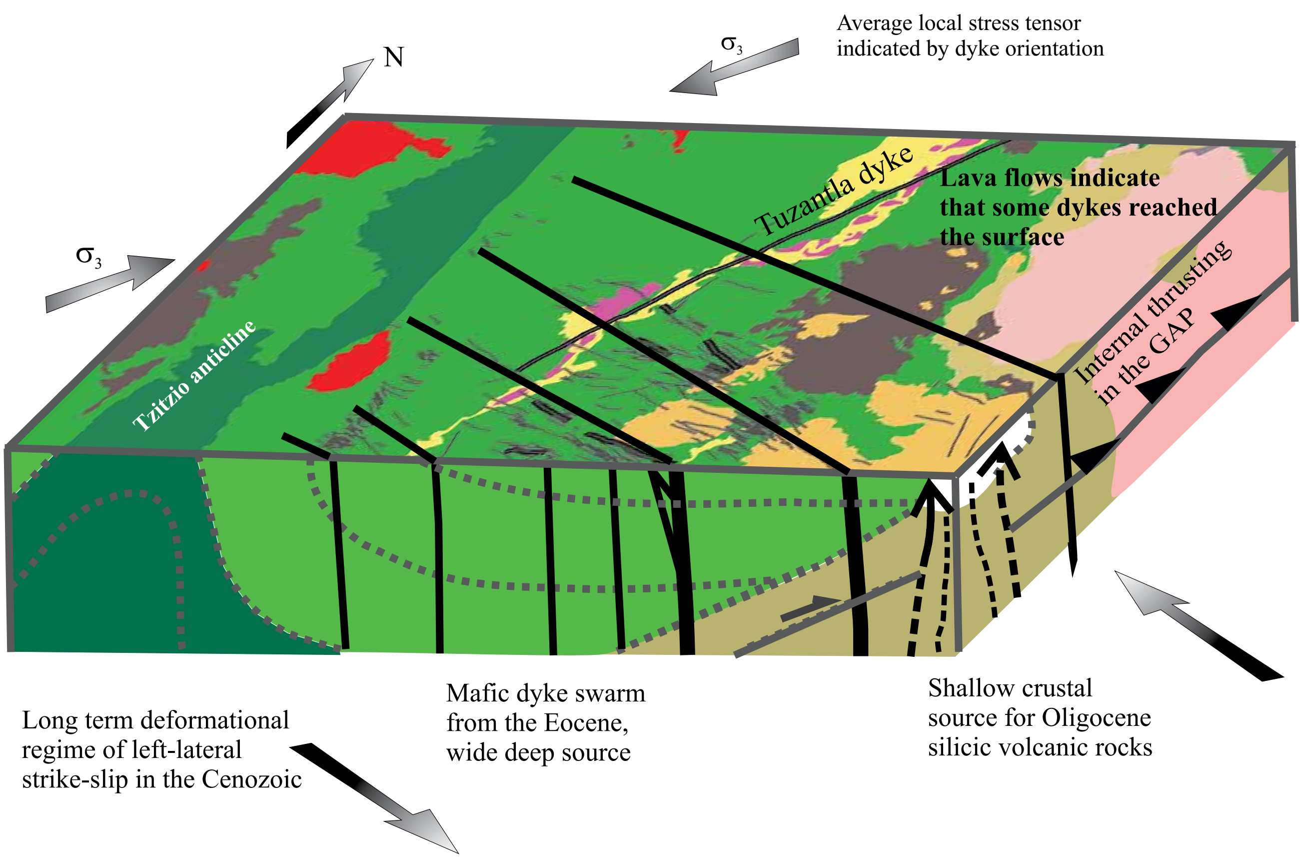

Figure 14. Schematic and somewhat speculative cartoon showing the conceptual model of the stress and strain directions that prevailed during the emplacement of dykes and the contemporaneous shear direction characteristic of southern Mexico. The proposed conceptual model implies that overpressured magma allowed the intrusion of the dykes along faults associated to the overall transtensional setting.

Another interesting pattern of propagation was evidenced by the morphology of dykes and their interaction with the host rock. The *en echelon* dyke segments may indicate shearing that generated the steps (offsets) between segments. Usually the generation of *en echelon* dykes is attributed to a rotation of the remote stress field that twists the planar fracture and segments it (Delaney and Pollard, 1981). The stress reorientation might occur as dykes reach the surface (Rubin, 1995), as can be expected for the shallow levels we are observing. However, in a few cases we have observed a lateral emplacement of magma even at the very shallow levels of observation, which indicates that other mechanisms are playing a role. Experimental work (Sommer, 1969) also showed that the formation of segments is due to the local adjustment of the crack plane to the changes in the direction of maximum principal stress. This observation implies the application of a shear deformation around an axis perpendicular to dyke walls. We assume that this can be attributed to a local change in the stress tensor, since it occurs occasionally in all the area of the Nanchititla dyke swarm. Furthermore, the steps developed in vertical sections are patterns observed in well-bedded strata (Gudmundsson, 2002) where it is clear that dyke passes through mechanically heterogeneous layers of scoria, tuff, or breccia and where the two parts of the dyke are connected by thin veins. These thin veins form in a layer promoting a lateral displacement of magma that eventually would form the offset between the segments. In our case, we observed four cases of dyke segmentation (Figure 5), in which adjacent dykes located at less than 5 meters in distance are not segmented, ruling out a subsequent layer displacement (faulting) as the cause of the offset. Our observations confirm the presence of dyke segmentation; however, it is unclear if the mechanical contrast of the stratigraphic layers is responsible for this behaviour of the dykes since segments are not observed connected through veins.

CONCLUSIONS

The excellent exposures of the Nanchititla dyke swarm of southern Mexico make it a natural laboratory to explore the evolution and propagation of dyke injection in the uppermost part of the continental crust. This dyke swarm is one of the few known cases in which a continental mafic dyke swarm is not directly related to extensional tectonics but to transtension (*e.g.*, Scarrow *et al.*, 2011). The dyke swarm appears related to the start of a large-scale and long-lived shear zone characterized by left-lateral displacement in the Early Cenozoic, where both the magmatic overpressure and the regional strain field controlled the propagation of dykes. Lateral migration of magma might have played an important role during propagation. Magma injection appears coeval with distributed strike-slip deformation of southern Mexico during the Eocene. The frequent presence of large phenocrysts indicates that some magma cooled below their liquidus temperatures and resided at depth before being injected. Probably, the magmas impregnated a wide zone of the lower crust and, toward the east were able to promote the formation of silicic magma chambers that eventually formed the Nanchititla Silicic Centre.

Field observations suggest that dykes were emplaced in a mechanically heterogeneous host rock. The mechanically weaker siltstone-bearing sequences allow a heterogeneous interaction with the overpressured magmatic fluid, whereas volcanic breccia and conglomerate behave in a stiffer manner, producing fractures parallel to dykes and brecciation.

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