Geological Society of America
Special Paper 422
2007

Geology and eruptive history of some active volcanoes of México

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ABSTRACT

Most of the largest volcanoes in México are located at the frontal part of the Trans-Mexican Volcanic Belt and in other isolated areas. This chapter considers some of these volcanoes: Colima, Nevado de Toluca, Popocatépetl, Pico de Orizaba (Citlaltépetl), and Tacaná. El Chichón volcano is also considered within this group because of its catastrophic eruption in 1982. The volcanic edifice of these volcanoes, or part of it, was constructed during the late Pleistocene or even during the Holocene: Colima 2500 yr ago, Pico de Orizaba (16,000 yr), Popocatépetl (23,000 yr), Tacaná (~26,000 yr), and Nevado de Toluca (>50,000). The modern cones of Colima, Popocatépetl, Pico de Orizaba, and Tacaná are built inside or beside the remains of older caldera structures left by the collapse of ancestral cones. Colima, Popocatépetl, and Pico de Orizaba represent the youngest volcanoes of nearly N-S volcanic chains. Despite the repetitive history of cone collapse of these volcanoes, only Pico de Orizaba has been subjected to hydrothermal alteration and slope stability studies crucial to understand future potential events of this nature.

The magmas that feed these volcanoes have a general chemical composition that varies from andesitic (Colima and Tacaná), andesitic-dacitic (Nevado de Toluca, Popocatépetl, and Pico de Orizaba) to trachyandesitic (Chichón). These magmas are the result of several magmatic processes that include partial melting of the mantle, crustal assimilation, magma mixing, and fractional crystallization. So far, we know very little about the deep processes that occurred between the upper mantle source and the lower crust. However, new data have been acquired on shallower processes between the upper crust and the surface. There is clear evidence that most of these magmas stagnated at shallow magma reservoirs prior to eruption; these depths vary from 3 to 4 km at Colima volcano, ~6 km at Nevado de Toluca, and 10–12 km at Chichón volcano.

Over the past 15 years, there has been a surge of studies dealing with the volcanic stratigraphy and eruptive history of these volcanoes. Up to the present, no efforts have achieved integration of the geological, geophysical, chemical, and petrological information to produce conceptual models of these volcanoes. Therefore, we still have to assume the size and location of the magma chambers, magma ascent paths, and time intervals prior to an eruption, in order to construct hazard maps that ultimately will establish permanent long-running monitoring systems. Today, only Colima and

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Macías, J.L., 2007, Geology and eruptive history of some active volcanoes of México, *in* Alaniz-Álvarez, S.A., and Nieto-Samaniego, Á.F., eds., Geology of México: Celebrating the Centenary of the Geological Society of México: Geological Society of America Special Paper 422, p. 183–232, doi: 10.1130/2007.2422(06). For permission to copy, contact editing@geosociety.org.

Popocatépetl have permanent monitoring networks, while Pico de Orizaba, Tacaná, and Chichón have a few seismic stations. Of these, Popocatépetl, Colima, and Pico de Orizaba have volcanic hazard maps that provide the basic information needed by the civil defense authorities to establish information programs for the population as well as evacuation plans in case of a future eruption.

Keywords: geology, eruptive chronology, active volcanoes, México.

INTRODUCTION

The first reports regarding volcanic activity in México appeared in the náhuatl codes; the volcanoes that deserved such attention were Pico de Orizaba (Citlaltépetl) and Popocatépetl. Their eruptions were shown in these codes as hills with a smoking top. The best record belongs to Popocatépetl, which means in náhuatl language "the smoking mountain." The Aztecs reported the occurrence of eruptions in 1363, 1509, 1512, and 1519–1528 (De la Cruz-Reyna et al., 1995). After the Spanish conquest, several volcanoes received more attention mostly focused on the activity of Popocatépetl volcano. However, it is only until the beginning of the nineteenth century that some scientific observations were carried out, such as those by Humboldt in 1804, Del Río in 1842, Del Castillo in 1870, and Sánchez in 1856 (in De la Cruz-Reyna et al., 1995). During this time several travelers, and among them some painters, produced detailed images of the volcanoes: Echeverría in 1793, Rugendas between 1831 and 1834, Baptiste in 1834, Egerton in 1834, Pieschel in 1856, Sattler in 1862, and White between 1862 and 1863. Their paintings provide information regarding the activity of several of the largest volcanoes in México, but they do not represent scientific studies aimed to understand their behavior. It was not until 1890 that two European travelers visited México and pointed out the distribution of the largest volcanoes (Felix and Lenk, 1890). In their report, they proposed that volcanoes like Popocatépetl, Cerro del Ajusco, Nevado de Toluca, Patzcuaro, Patambán, and La Bufa de Mascota (Bufa de Real Alto) were genetically linked to the presence of a single fissure. They also mentioned the existence of aligned volcanoes in secondary fissures, such as Pico de Orizaba–Cofre de Perote, el Telapón, Tlamacas, Iztaccíhuatl, Popocatépetl, and Ceboruco–Cerro de la Bufa. Without a specific purpose except to understand the nature of the Mexican volcanoes, the observations of Felix and Lenk (1890) produced great skepticism amongst the small Mexican geological community who began to investigate the origin, age and distribution of the volcanoes (Ordoñez, 1894). At that time, only general geological studies had been performed on some volcanoes, and some details of the relative age of certain volcanic chains were known. México had a reduced number of geologists who were mostly appointed to the National Geological Survey and to the Geological Commission of México. Ordoñez (1894), then a member of the Geological Commission and one of the nation's most prestigious geologists, reviewed the work by Felix and Lenk (1890) and denied the existence of a large fracture that produced all these volcanoes. He discussed the differences between the truncated crater of Nevado de Toluca and the Quaternary eruptions of Popocatépetl, the Ajusco volcano that did not have a crater, and the contemporaneity and common genesis of the volcanoes. Analyzing the existence of N-S faults, Ordoñez considered that the pyroxene-andesites of Iztaccíhuatl volcano and others in México were related to volcanoes that were destroyed in the past—in other words, old volcanoes, as opposed to the hornblende-andesites of Popocatépetl, which had a young crater and notorious activity. In this way, Ordoñez rejected the existence of secondary N-S fissures. Despite of this erroneous conclusion, Ordoñez demonstrated that he had a broad knowledge of the Mexican volcanoes, as reflected in his studies of Iztaccíhuatl (Ordoñez, 1894), Popocatépetl (Aguilera and Ordoñez, 1895), Colima and Ceboruco (Ordoñez, 1898), Cofre de Perote (Ordoñez, 1905), and Nevado de Toluca (Ordoñez, 1902). In these studies, he defined the morphologic features of the volcanic edifices, the petrographic nature of the rocks, and stratigraphic descriptions of the deposits. His work established the basic frame for new geological studies of the Mexican volcanoes, which started at the end of the nineteenth century and the beginning of the twentieth century. The first meteorological and volcanological observatory in México was founded in 1893 in Zapotlán, Jalisco, by priests Arreola and Diaz, who performed systematic observations of Colima volcano between 1893 and 1905. In 1895, Arreola founded the Colima Volcanological Observatory (Díaz, 1906), a surprising fact considering that the first volcanological observatory in the world was established at Vesubio volcano (Italy) by Palmieri only 54 years earlier, after a series of eruptions.

The first decade of the twentieth century had a large impact in the geological community worldwide, due to the occurrence of several volcanic eruptions in Latin America. On 8 May 1902, Mount Pelée on the island of Martinique erupted, producing a pyroclastic surge that completely destroyed the city of St. Pierre, killing more than 25,000 people (Perret, 1937). A day before, on 7 May 1902, Soufrière volcano devastated the island of St. Vincent, with a total death toll of 2000 people (Anderson and Flett, 1903). Finally, in October of the same year, Santa Maria volcano, Guatemala, awoke violently, producing a Plinian eruption that covered the NW portion of Guatemala with pumice and sent ash into central México. This eruption generated 12 km^3 of magma, which represents one of the largest eruptions in the twentieth century (Williams and Self, 1983). The catastrophic eruption of Mount Pelée attracted the attention of geologists worldwide, who studied it and subsequent eruptions in detail and established the Volcanological Observatory of Morne des Cadets in 1903.

These eruptions opened a new course in the study of active volcanoes in México, such as those of Böse (1903) about Tacaná volcano in the State of Chiapas. Böse concluded that Tacaná volcano consisted of three peaks that represented the remains of old edifices. In addition, he performed detailed observations of the summit, where he described small craters produced by recent eruptions. Waitz (1909) concluded that Nevado de Toluca volcano represented the ruins of an andesitic stratovolcano, with an extinct central dome affected by glacier activity, due to the presence of moraine deposits.

During the second decade of the twentieth century, two important volcanic eruptions took place in México that influenced the development of volcanological studies. These events were the 1913 eruption of Colima volcano and the 1913–1927 eruption of Popocatépetl. Despite the fact that these eruptions occurred during the troubled times of the Mexican Revolution, both attracted the attention of national and international geologists. This was followed by two decades of relative quietness, but this ended on 20 February 1943, when there was news of the birth of a volcano in a field corn of the State of Michoacán, close to San Juan Parícutin. This volcano, called Parícutin, reached an elevation of 424 m and buried the villages of Parícutin and San Juan Parangaricutiro (Flores-Covarrubias, 1945). This eruption was closely followed by volcanologists (Wilcox, 1954; González-Reyna, 1956; Segerstrom, 1956), and Parícutin became the most studied volcano in México (Luhr and Simkin, 1993). Today, Parícutin forms part of education programs at the elementary level and is a classic example of the birth of a volcano worldwide. In 1956, a spectacular eruption occurred on San Benedict Island in the Revillagigego archipelago, giving rise to the Bárcena volcano. Because of its remote location, this eruption was studied by only a small group of volcanologists (Richards, 1959, 1965). In 1962, the crater of Colima volcano, formed in 1913, was filled to the brim, beginning the emission of lava and pyroclastic flows on its flanks, with several eruptions occurring in 1962, 1976, 1981, 1987, 1991, 1994, 1998–2000, and 2002–2005.

The event that accelerated volcanological studies in México was the 28 March 1982 catastrophic eruption of El Chichón volcano in the State of Chiapas (Sigurdsson et al., 1984). The reactivation of this practically unknown volcano took place after 550 years of quiescence (Tilling et al., 1984; Macías et al., 2003), taking scientists by surprise. This eruption had a death toll of 2000 people, destroyed nine villages, and caused important global effects in the planet with the emission of 7 Mt of SO_2 into the atmosphere and the reduction of the planet temperature of ~0.5 °C during several months (Espíndola et al., 2002). The El Chichón eruption still represents the worst volcanic disaster in México in historic times. As a consequence of this eruption and the 1985 earthquake that devastated México City, the National Civil Protection System was created. During the late eighties, the volcanic activity continued with small events, such as the 1986 phreatic eruption of Tacaná volcano that ended with the emission of a fumarole. This event warned the state authorities to begin monitoring Tacaná volcano. During the 1990s, the small submarine eruption of Everman volcano in the Revillagigedo archipelago occurred (Siebe et al., 1995a). Another important event was the reactivation of Popocatépetl volcano occurred on 21 December 1994; it not only represented a breakthrough in volcanological studies, but for hazard mitigation studies as well. Immediately after the crisis began, a scientific committee was formed to evaluate the volcanic crisis at Popocatépetl. This committee deemed it pertinent to construct a volcanic hazards map for civil protection authorities. It delineated emergency plans that included evacuation routes, meeting points, shelters, etc. The Popocatépetl hazards map was published in February 1995 (Macías et al., 1995). This map was followed by the creation of the Colima volcano (Martín Del Pozzo et al., 1995; Navarro et al., 2003) and Pico de Orizaba (Sheridan et al., 2002) hazard maps.

LOCATION OF THE ACTIVE VOLCANOES IN MÉXICO

The highest concentration of volcanoes in México is located in the Trans-Mexican Volcanic Belt, where a large variety of volcanic landforms occurred as monogenetic volcanic fields, majestic stratovolcanoes with elevations higher than 4000 m above sea level, shield and composite volcanoes, calderas, fissural lava flows and domes (Fig. 1). Inside the Trans-Mexican Volcanic Belt there are several N-S to NE-SW volcanic chains made of stratovolcanoes and composite volcanoes, where the volcanic activity has migrated during the past 2 m.y., toward the frontal part of the volcanic arc. In other words, the active volcanoes are located in the southward end of this volcanic chain. These volcanic chains are the Cántaro–Nevado de Colima–Colima, Tláloc–Telapón–Iztaccíhuatl–Popocatépetl, and Cofre de Perote–Las Cumbres–Pico de Orizaba–Sierra Negra. In this work, some of the large active volcanoes in México are considered, such as Colima, Nevado de Toluca, Popocatépetl, Pico de Orizaba, and Tacaná. Due to its 1982 catastrophic eruption, El Chichón is also considered within this group. In the following section, there is a description of each volcano, a compilation of the available information, and then an analysis and reflection of the advances and limitations of such studies, as well as another studies yet to be accomplished.

COLIMA VOLCANO

Colima volcano or Fuego de Colima (19°30′45″, 103°37′) has an elevation of 3860 m above sea level and is the eighth highest peak in México (Fig. 2). The volcano is located 100 km south of the city of Guadalajara and 30 km north of the city of Colima. Colima volcano belongs to an N-S volcanic chain formed by the Cántaro, Nevado de Colima, and Colima volcanoes.

Previous Studies

The first geological studies of Colima volcano were carried out by Waitz (1906, 1915, 1935). Waitz made general observations of the volcano and described the generation of pyroclastic

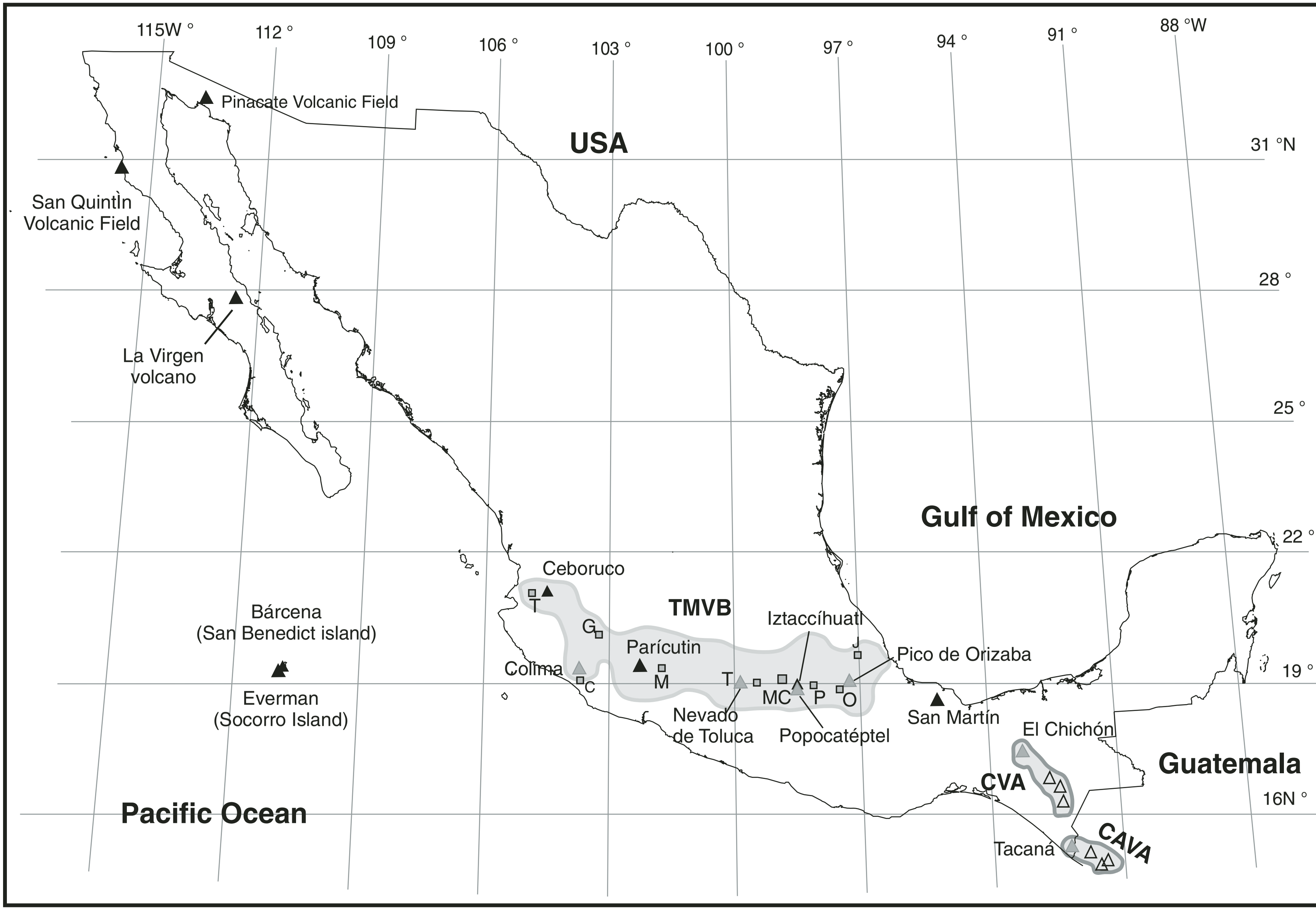


Figure 1. Location of main volcanoes in México (black triangles) that are grouped along the 19–20°N parallel of latitude forming the Trans-Mexican Volcanic Belt (TMVB). There are other regions of isolated volcanism, such as the Revillagigedo archipelago, the Tres Vírgenes Volcanic Complex, Los Tuxtlas Volcanic Complex, and the Chiapanecan Volcanic Arc. The volcanoes treated in this work appear as gray triangles. CVA—Chiapanecan Volcanic Arc; CAVA—Central America Volcanic Arc; C—Colima; G—Guadalajara; J—Jalapa; M—Morelia; MC—México City; O—Orizaba; P—Puebla; T—Tepic.

Figure 2. Panoramic view of the southern flank of Colima volcano taken in December 1998. Visible are some blocks collapsed from the summit that roll down the steep slopes of the volcano.

flows during the 1913 eruption. From that time until the end of the 1950s, Colima volcano's activity remained restricted to the crater's interior. In 1962, the crater was completely filled with lava, causing the emission of lava flows on the volcano flanks, attracting the attention of volcanologists who studied the morphology and some general features of the volcanic complex (Mooser, 1961), its Merapi type pyroclastic flows (Thorpe et al., 1977), and the evolution of the volcano (Demant, 1979). The 1980s were the starting point of the modern studies of Colima volcano. After the 1981 eruption, Medina-Martínez (1983) presented the first analysis of eruptive recurrence of the volcano during the past 400 yr. In addition, the first chemical and petrological studies of the volcano and adventitious cones were completed (Luhr and Carmichael, 1980, 1981, 1982, 1990a) as well as a study of the collapse of the volcanic edifice (Robin et al., 1987). This latter study created an enormous interest in the volcano and led to a surge of new studies and discussion over the subsequent years (Luhr and Prestegaard, 1988; Stoopes and Sheridan, 1992; Komorowski et al., 1997; Capra and Macías, 2002; Cortés, 2002). The 1990s were distinguished by the generation of several small eruptions at Colima that produced lava and pyroclastic flows. These eruptions were studied by different specialists, from seismic studies (Núñez-Cornu et al., 1994; Jiménez et al., 1995; Domínguez et al., 2001; Zobin et al., 2002), petrology (Robin et al., 1990; Robin and Potrel, 1993; Connor et al., 1993; Luhr, 2002; Macías et al., 1993; Mora et al., 2002; Valdez-Moreno et al., 2006), gas chemistry (Taran et al., 2002), and stratigraphy (Martín Del Pozzo et al., 1987; Rodríguez-Elizarrarás et al., 1991; Rodríguez-Elizarrarás, 1995; Navarro-Ochoa et al., 2002; Saucedo et al., 2002; Saucedo et al., 2004a). During the past 25 years, several geological maps of Colima volcano have been completed (Demant, 1979; Luhr and Carmichael, 1990b; Rodríguez-Elizarrarás, 1991; Cortés et al., 2005), as well as several types of hazards maps (Sheridan and Macías, 1995; Martín Del Pozzo et al., 1995; Navarro et al., 2003; Saucedo et al., 2004b).

Evolution of the Volcanic Chain

The NW part of the Trans-Mexican Volcanic Belt is subjected to the subduction of the Rivera plate underneath the North America plate and to the triple point junction formed by the Tepic-Zacoalco rift to the NW, the Chapala rift to the E, and the Colima graben to the S (Luhr et al., 1985; Garduño and Tibaldi, 1990) (Fig. 3). The first two rifts bound to the N and to the E of the so-called Jalisco block and are considered to be old cortical structures reactivated by forces acting at the plate boundaries (Rosas-Elguera et al., 1996). According to Cortés et al. (2005), the Colima Volcanic Complex is built upon the lower Cretaceous andesites and volcaniclastic deposits of the Tecalitlán Formation, sandstones and claystones of the Encino Formation, massive limestones of the Tepames Formation, red beds of the upper Cretaceous Coquimatlan Formation, Cretaceous intrusive rocks, and a Tertiary volcanic sequence made of basaltic and andesitic lava flows, dacitic volcanic breccias, and ignimbrites (Fig. 4).

The Quaternary volcanic activity in the graben began ca. 1.6 Ma with formation of the Cántaro stratovolcano (Allan, 1986). Allan and Carmichael (1984) reported K-Ar ages for this volcano that vary from 1.66 ± 0.24, 1.52 ± 0.20, and 1.33 ± 0.20 Ma. Cántaro volcano consists of andesitic lava flows followed by dacitic domes (Luhr and Carmichael, 1990b). The activity of Cántaro ended ca. 1.0 Ma.

Nevado de Colima Volcano

Afterward, the volcanic activity migrated ~15 km to the south to form the ancient Nevado de Colima volcano, which had a complex eruptive history composed of either two eruptive phases (Mooser, 1961), four phases (Robin et al., 1987), or six eruptive periods (Cortés et al., 2005). According to the latter authors, these eruptive periods were influenced by the activity of the Tamazula fault (Garduño et al., 1998); three of them are associated with the failure of the volcano and the generation of debris avalanche deposits. These eruptive periods are described below:

1. During the first eruptive period, ca. 53 Ma, a volcanic edifice 25 km in diameter was formed (Robin et al., 1987). This stratovolcano was composed of andesitic lavas, pyroclastic flows, and fallout deposits reaching a volume higher than 300 km^3.

2. The formation of a second edifice smaller than the first made of andesitic lava and pyroclastic flow and fall deposits of unknown age.

3. The construction of a third edifice composed of andesitic lava and pyroclastic flow and fall deposits. This edifice is associated with a 4-km-wide semicircular caldera open to the SE, produced by the collapse of the flank that caused a debris avalanche deposit exposed to the NE of the El Platanar village.

4. The activity continued with the building of a new volcanic edifice with the emission of 17 km long andesitic lava flows,

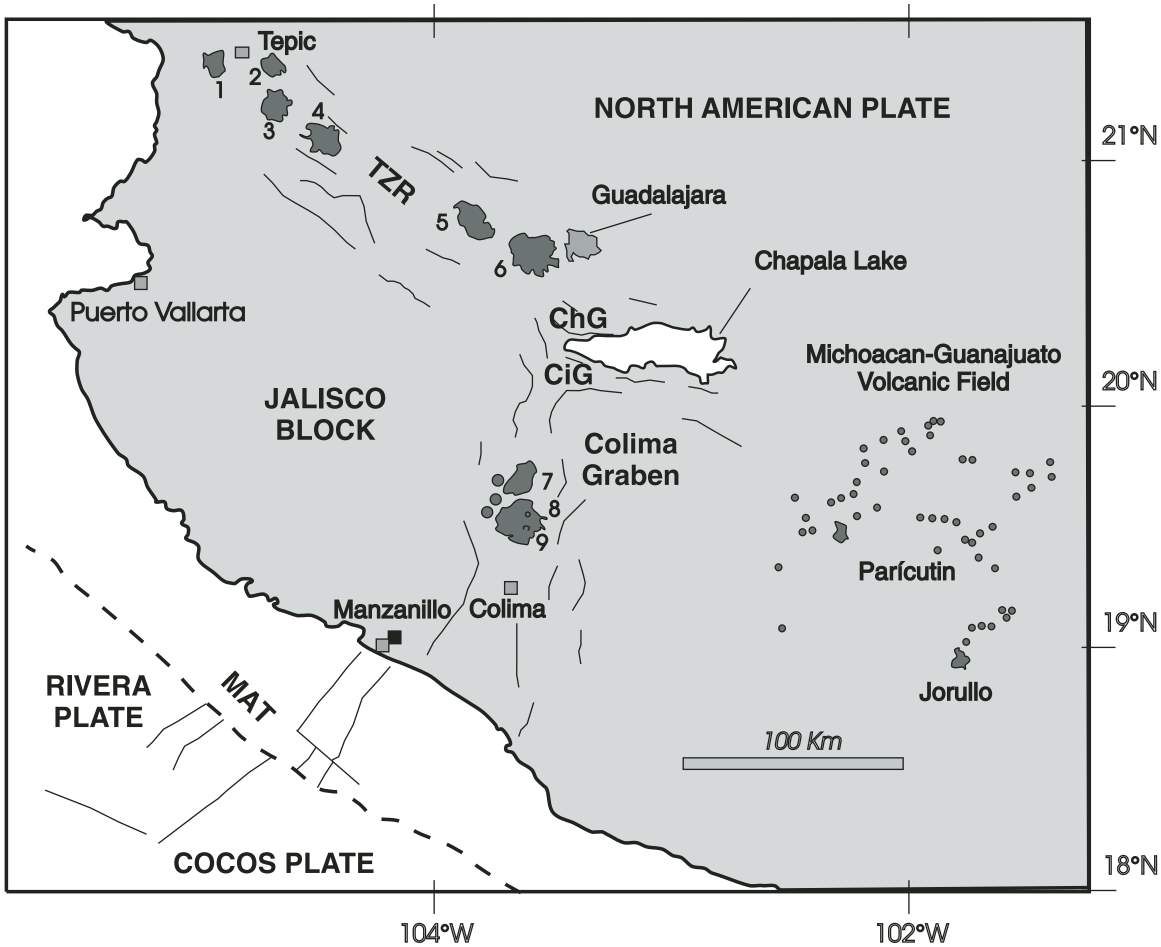


Figure 3. Tectonic setting of the western portion of the Trans-Mexican Volcanic Belt, after DeMets and Stein (1991), Lange and Carmichael (1991), and Luhr et al. (1989). The stratovolcanoes are 1—San Juan; 2—Sangangúey; 3—Tepetiltic; 4—Ceboruco; 5—Tequila; 6—La Primavera caldera. Colima Volcanic Complex: 7—Cántaro; 8—Nevado de Colima; 9—Colima volcano. ChG—Chapala Graben; CiG—Citala Graben; MAT—Middle America Trench (discontinuous line); TZR—Tepic-Zacoalco Rift. Thin lines represent fractures.

pyroclastic flows, and fallouts. Some of these lava flows have an age of 0.35 Ma (Robin and Boudal, 1987). The activity of the volcano ended with the lateral collapse of the edifice toward the SE, originating a second debris avalanche deposit exposed up to 25 km to the SE of the crater at the Beltrán and Tuxpan-Naranjo gullies. The precise age of the collapse is still unknown.

5. Renewed activity began at the interior of the caldera with the emplacement of thick lava flows and pyroclastic deposits exposed at the Atenquique gully; these deposits built a fifth volcanic edifice whose activity ended with an explosive event that produced block-and-ash flows traveling 17 km to the ESE from the crater to the Atenquique area and leaving a semicircular caldera opened to the ENE (Cortés et al., 2005). These deposits discordantly overlie the fluviatile deposits of the Atenquique Formation (Mooser, 1961) that Robin and Boudal (1987) dated at 0.38 and 0.26 Ma. However, this caldera has been associated by other authors with the gravitational collapse of Nevado de Colima, which occurred 18,500 yr B.P. and generated a third debris avalanche deposit (Robin et al., 1987; Stoopes and Sheridan, 1992; Capra, 2000; Capra and Macías, 2002). This debris avalanche deposit caused an enormous interest in the scientific community since Stoopes and Sheridan (1992) concluded that this avalanche had traveled along the Tuxpan-Naranjo and Salado rivers all the way to the Pacific Coast across a distance of 120 km from the summit, at that time the largest described in the world. However, Capra (2000), Pulgarín et al. (2001), and Capra and Macías (2002) stated that this event, in fact, began with the flank failure of Nevado de Colima that produced a debris avalanche that traveled 25 km up to the Naranjo River. The deposit dammed the river, forming a temporary lake that, after a few days, broke to produce a gigantic debris flow, which moved along the Tuxpan-Naranjo rivers up to the Pacific Ocean.

6. This event was followed by a period of volcanic quiescence of Nevado de Colima, which allowed the development of

soils and occasional explosive eruptions, forming extended yellow ash flows with pumice (Robin and Boudal, 1987). One of these deposits was dated at 17,960 yr B.P. at the NW edge of the Quesería village, whose age is very close to the younger debris avalanche of the volcano. It has been assumed that these may be associated events or closely spaced in time. In addition, there was a series of Plinian eruptions that occurred between 8000 and 2000 yr B.P. that emplaced fallouts and pyroclastic surges at the caldera interior (Navarro-Ochoa and Luhr, 2000; Luhr and Navarro-Ochoa, 2002). The eruptive activity of Nevado de Colima came to an end with andesitic lava flows contained inside the caldera walls and the emplacement of the El Picacho dome, which is the present summit of the volcano.

Paleofuego Volcano

At the same time as the last stages of activity of Nevado de Colima, Paleofuego began its construction 5 km south of the ancient cone of Colima volcano (Robin and Boudal, 1987) (Fig. 4). This volcanic edifice is now represented by a 5-km-wide caldera open to the south. The projection of the caldera walls suggests that the volcano had an approximate elevation of 4100 m above sea level (masl) (Luhr and Prestegaard, 1988). The 300-m-thick northern walls are made of andesitic lava flows alternated with block-and-ash flow deposits (Cortés et al., 2005). These lava flows moved 17.5 km to the SW and 31 km to the SE. A sequence composed of a pyroclastic flow, lahar, and lacustrine beds was dated at 38,400 yr B.P. by Komorowski et al. (1993), who associated it with Paleofuego volcano. Therefore, this age represents the minimum age of formation of Paleofuego. Waitz (1906) was the first author to describe the Paleofuego caldera as a maar; Demant (1979) associated its formation with a series of cyclic eruptions of scoria and ash; Robin et al. (1987) and Luhr and Prestegaard (1988) interpreted it as the remains of a structure from a Mount St. Helens–type collapse. Luhr and Prestegaard (1988) estimated an area of 1550 km^2 for the debris avalanche deposit, located to the south of the volcano, and dated charcoal material at its base at 4280 ± 110 yr B.P. However, Robin et al. (1987) assigned an older age of 9370 ± 400 yr B.P. to the deposit and considered it to be composed of several deposits. Later, this deposit was mapped by Komorowski et al. (1997), who concluded that the deposit represented the youngest debris avalanche of the volcano, which traveled 30 km to the south, covered a surface of ~1200 km^2 and occurred 2500 yr B.P. More recently, Cortés et al. (2005) reported that Paleofuego volcano collapsed at least five times during its eruptive history, with deposits covering an area of 5000 km^2 (Fig. 4). The following summary of events and deposits uses the abbreviations presented by Cortés et al. (2005):

The first debris avalanche (CVP3) crops out 40 km south of the volcano, in the vicinity of Coquimatlán. This deposit develops a series of small hummocks aligned in the flow direction. The deposit covers a surface of ~445.5 km^2 and has an average thickness of 20 m and a volume of 8.9 km^3. It is covered by a sequence of fluviatile and lahar deposits, divided by a paleosol dated at 16650 ± 135 yr B.P.

The second debris avalanche deposit (CVP4) outcrops near the town of Mazatán and is located to the west of the volcano, atop the Cerro Grande limestones. The deposit is covered by a lacustrine sequence dated at 6390 and 7380 yr B.P. (Komorowski et al., 1997).

The third deposit (CVP5) covers ~586 km^2 and has a volume of ~30 km^3. Outcrops from 0 to 15 km from the crater consist of block facies (hummocks) with close depressions forming lakes (El Jabalí, Carrizalillos, etc.). Here the blocks consist of red and gray lava with jigsaw-fit structure and scarce matrix. From 15 to 30 km from the volcano, the deposit developed a matrix facies with a morphology of smooth hills. This facies consists of angular blocks (up to 2 m in diameter) set in a poorly consolidated fine ash matrix. Organic material found inside the deposit was dated at 6990 ± 130 yr B.P. (Cortés and Navarro-Ochoa, 1992).

The fourth debris avalanche deposit (CVP6) extends up to 20 km to the SW, from the present Paleofuego caldera wall, and covers ~40 km^2. This deposit consists of megablocks assimilated from different volcanic and fluvio-lacustrine deposits (Cortés, 2002) and was dated at 3600 yr B.P. (Komorowski et al., 1997). The fifth debris avalanche is represented by the youngest deposit dated at 2500 yr B.P. by Kmorowski et al. (1997).

Colima Volcano

After the last collapse of Paleofuego volcano 2500 yr ago, the volcanic activity migrated to the south inside the caldera floor of Paleofuego. This activity formed the modern Colima volcano, which has an approximate volume of 10 km^3 and has grown at a rate of 0.002 km^3/yr (Luhr and Carmichael, 1990a, 1990b). This stratovolcano is composed of an interlayering of andesitic lava flows and pyroclastic deposits (fall, surge and flow). One of the main features of Colima volcano is the generation of pyroclastic flow deposits that have reached 15 km from the summit, like those produced during the 1913 eruption. During the past 400 yr, Colima volcano has experienced ~43 eruptions, which places it among the most active volcanoes in North America (De la Cruz-Reyna, 1993; Saucedo et al., 2004b).

The most detailed studies summarizing the eruptive activity of Colima volcano are those by Medina-Martínez (1983), De la Cruz-Reyna (1993), Saucedo and Macías (1999), and Bretón et al. (2002). The oldest eruptions occurred in the sixteenth and seventeenth centuries (Tello, 1651), including the 13 December 1606 (Arreola, 1915) and the 15 April 1611 (Bárcena, 1887) eruptions, as well as those that took place in 1690 (De la Cruz-Reyna, 1993) and in 1771 (Bárcena, 1887), and many other minor eruptions. A detailed historical record began with the 15 February 1818 eruption (Sartorius, 1869), which destroyed a lava dome (Dollfus and Monserrat, 1867) and dispersed ash and scoria to the cities of Guadalajara, Zacatecas, Guanajuato, San Luis Potosí, and México (Bárcena, 1887; Arreola, 1915). After the 1818 eruption, Colima volcano had a 500-m-wide, funnel-shaped crater with 50–230-m-deep vertical walls. On 12 July 1869, the formation of the parasitic vent "El Volcancito" began; it ended in 1872 (Sartorius, 1869; Bárcena, 1887). Orozco et al. (1869) reported

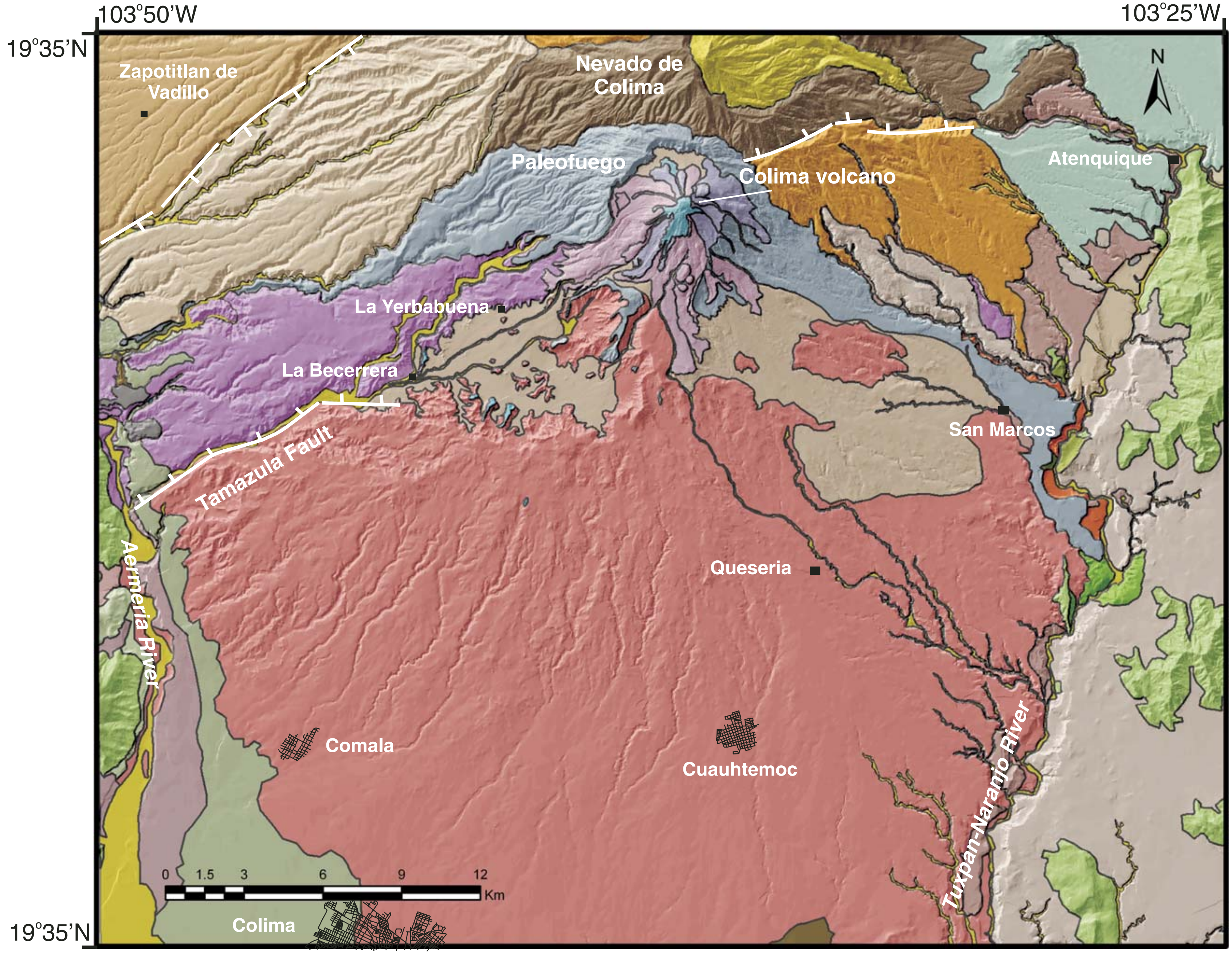


Figure 4 (*legend on following page*). Simplified geologic map of the southern portion of the Colima Volcanic Complex (taken from Cortés et al., 2005) superimposed on a digital elevation model. For the sake of simplicity, only the mapped units of Colima volcano and a portion of Nevado de Colima are labeled.

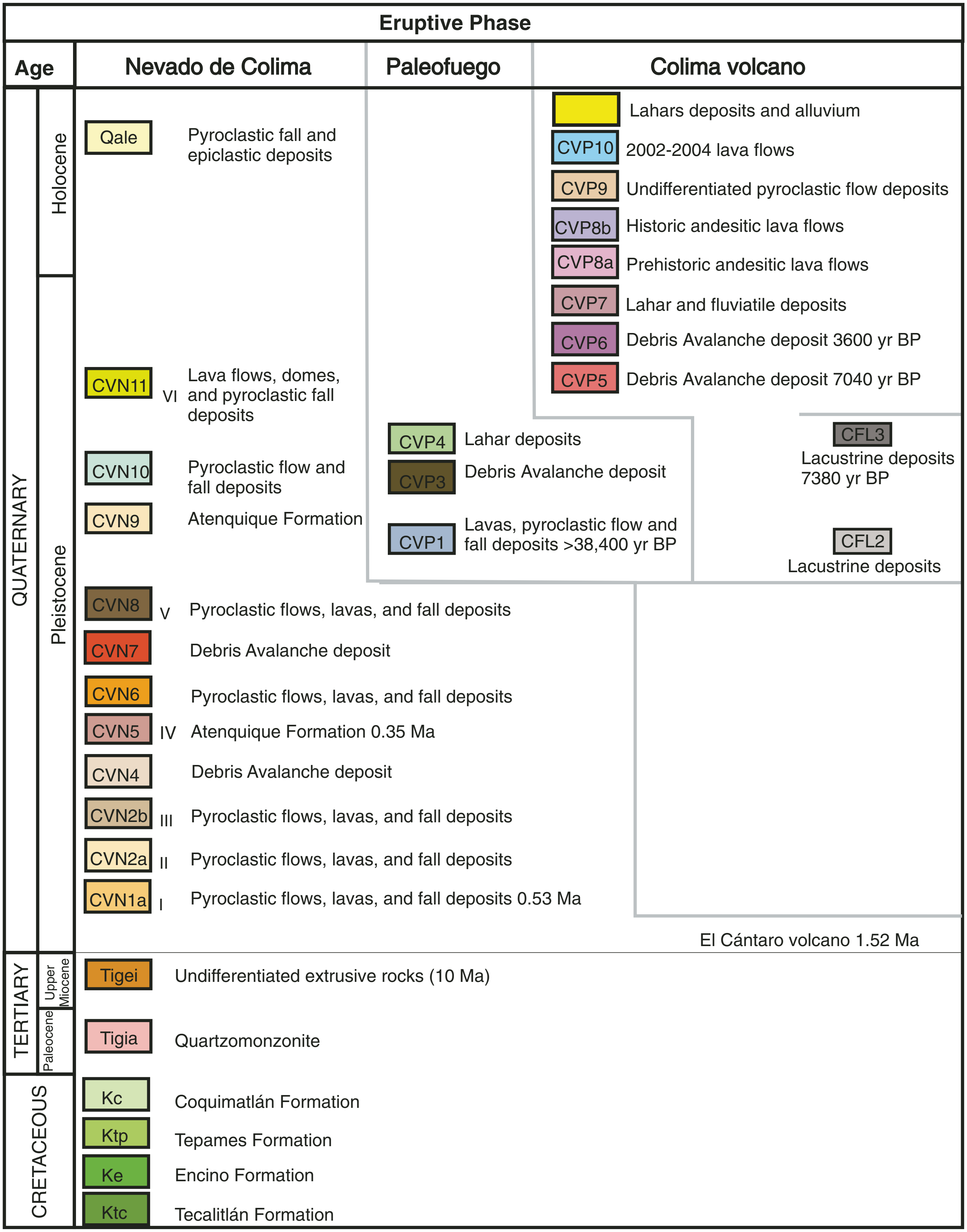


Figure 4 (*legend*).

that "El Volcancito" flared up at an elevation of 3200 m, reaching 300 m high for a maximum elevation of 3500 m. Between 1893 and 1903, the priest Arreola in Colima, Colima, and his colleague Castellanos in Zapotlán, Jalisco (Cuidad Guzmán today), started systematic observations of the volcano with the installation of volcanological observatories. The results of these observations were published in the Central Meterological Observatory bulletin in México City (Arreola, 1915). These observations were very important because the volcano was headed toward a major eruption, which took place in 1913. Prior to the eruption, the volcano had a summit crater filled by a dome.

The eruption began on 17 January 1913 (Ortíz-Santos, 1944) with a series of explosions that generated dense ash columns rich in vapor. On 20 January, the eruption continued with the formation of a 21-km-high Plinian column that was dispersed to the northeast by stratospheric winds (Saucedo, 1997). The column deposited 15 cm of pumice at Zapotiltán, Jalisco, and 4 cm at La Barca, Jalisco, according to the *El Impacial* journal published on 21 January. The ash of this column reached the city of Saltillo, located 700 km northeast of the volcano. The column collapsed, producing 15-km-long pyroclastic flows that descended to the southern flank of the volcano (Waitz, 1915). As a result of the eruption, the volcano summit had lost 100 m and a new 400-m-wide jogged crater was formed. The depth of this crater oscillated between 50 and 100 m (Waitz, 1935). In 1958, this crater was almost completely occupied by a blocky lava dome (Mooser, 1961) that, between 1961 and 1962, started to spill over the northern flank of the crater, forming a lava flow that reached the location known as El Playón. The 1975–1976 eruptions generated lava flows and, for the first time since the 1913 eruption, some pyroclastic flows (Thorpe et al., 1977). Since this event, Colima has emitted lava domes, lava flows, and different types of pyroclastic flows. Pyroclastic flows have been produced by the gravitational collapse of the external parts of the summit dome such as those that occurred in 1982 (Luhr and Carmichael, 1990b) and 1991 (Rodríguez-Elizarrarás et al., 1991; Saucedo et al., 2004a), by the collapse of the frontal parts of advancing lava flows as in 1998 (Saucedo et al., 2002), by the collapse of low-altitude ash columns in 1991 (Sauceo et al., 2002), and by strong explosions produced at the crater in 2005 (Macías et al., 2006).

NEVADO DE TOLUCA VOLCANO

Nevado de Toluca volcano (99°45′W, 19°09′N) is located 23 km to the SW of the city of Toluca (Fig. 5). Its highest peak has an elevation of 4680 masl (the fourth highest mountain in México). Nevado de Toluca is also known as "Xinantécatl" (naked man in náhuatl language), although García-Martínez (2000) concluded that the proper náhuatl name is "Chicnauhtécatl" meaning "nine hills." The volcanic edifice is built upon a volcano-sedimentary metamorphic basement of Late Jurassic–Early Cretaceous age (Bonet, 1971; Bloomfield et al., 1977; Campa et al., 1974; Demant, 1981), Eocene rhyolitic ignimbrites, and Miocene andesitic lava flows (García-Palomo et al., 2002) (Fig. 6).

Nevado de Toluca is a composite volcano of late Pleistocene-Holocene age, made up of calc-alkaline andesites and dacites (Bloomfield and Valastro, 1974; Cantagrel et al., 1981; García-Palomo et al., 2002). The Nevado de Toluca crater has the remains of two ancient scars, located on the SE and NE flanks of the volcano, that are related to the partial collapse of the edifice. The northern flank of Nevado de Toluca has a relative elevation of 2015 m with respect to the Lerma river basin, and its southern flank has a relative elevation of 2900 m with respect to the Ixtapan de la Sal village (Fig. 5). The crater of Nevado de Toluca is truncated (Fig. 7A) and has a 2 × 1.5-km-wide elliptical crater with its main axis aligned to an E-W direction and opened to the east. The craters interior holds a dacitic central dome called the navel "El Ombligo" that separates the Sun and Moon lakes whose surfaces stand at an elevation of 4200 m (Fig. 7B). The water of these two lakes contains diatoms (Caballero, 1996), and has alkaline composition (Armienta et al., 2000). Prehispanic obsidian blades and pottery shards are found dispersed on the surface of the Ombligo dome and at the bottom of the lakes, since these sites were used first by the Matlazincas, followed by the Aztec people to perform religious ceremonies (Quezada-Ramírez, 1972). The surface of Nevado de Toluca was carved by glacier activity during late Pleistocene-Holocene time, as evidenced by moraine deposits and rock glaciers (Heine, 1976, 1988; Vázquez-Selem and Heine, 2004).

Previous Studies

The first studies of Nevado de Toluca described morphological and petrographic aspects of the volcano (Ordoñez, 1902; Otis, 1902; Hovey, 1907; Flores, 1906; Waitz, 1909). After that, the volcano remained unstudied until the 1970s, when the first geological and volcanological studies were accomplished (Bloomfield and Valastro, 1974; Bloomfield and Valastro, 1977; Bloomfield et al., 1977). These authors described two Plinian eruptions that produced the Lower Toluca Pumice, dated at ca. 24,000 yr B.P., and the Upper Toluca Pumice, dated at 11,600 yr B.P. In addition, during these studies, the authors established part of the modern eruptive history of Nevado de Toluca, as well as the calc-alkaline nature of its products (Whitford and Bloomfield, 1977). In 1982, Cantagrel et al. (1981) divided the evolution of the volcano in two main stages. The oldest stage was composed of andesitic lava flows that form the ancient edifice dated with the K-Ar method at 1.60 ± 0.12 and 1.23 ± 0.15 Ma. The youngest stage was composed of a complex sequence of volcaniclastic deposits that in the southern flank of the volcano had a minimum thickness of 100 m and an age of 100,000 yr. According to these authors, the activity between these two stages was mainly volcaniclastic.

During the 1990s, a new surge of studies at Nevado de Toluca began to understand the structural setting of the volcano (García-Palomo et al., 2000), the geology and eruptive history (Macías et al., 1997a; García-Palomo et al., 2002), the deposits produced during Plinian eruptions (Arce et al., 2003; Arce et al., 2005b; Capra et al., 2006), the collapse events (Capra

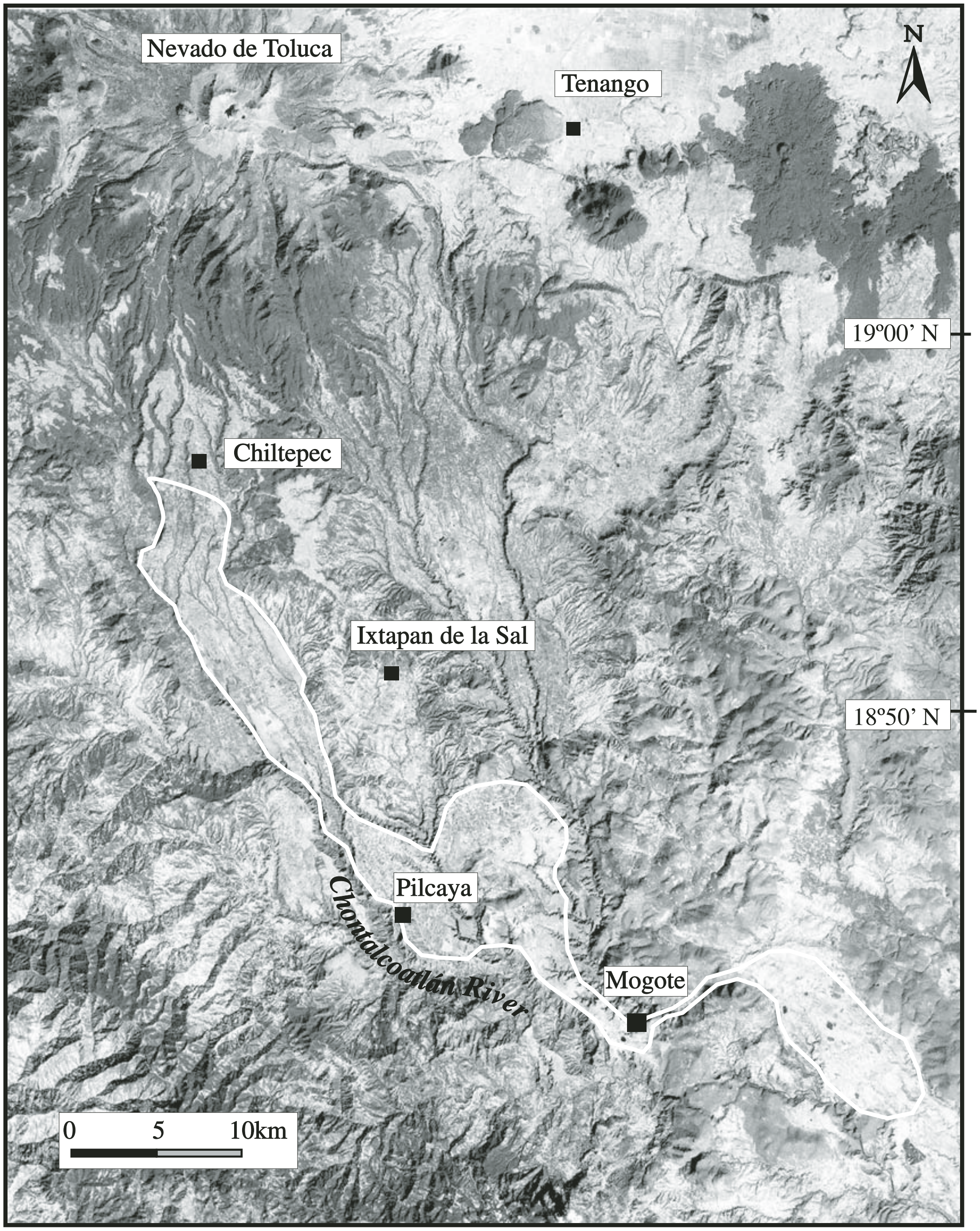


Figure 5. LANDSAT image showing Nevado de Toluca and the maximum extent of Pilcaya debris flow and the Mogote lahar deposits (Capra and Macías, 2002).

and Macías 2000), geomorphological aspects (Norini et al., 2004), the paleosols preserved between the volcanic deposits (Sedov et al., 2001, 2003; Solleiro-Rebolledo et al., 2004), and paleoenvironmental deposits in the Upper Lerma Basin (Metcalfe et al., 1991; Newton and Metcalfe, 1999; Caballero et al., 2001, 2002; Lozano-García et al., 2005).

Evolution of Nevado de Toluca

Nevado de Toluca was built upon the intersection of three fault systems with NW-SE, NE-SW, and E-W orientations (García-Palomo et al., 2000). This structural geometry favored the formation of coalescent pyroclastic fans, with smooth mor-

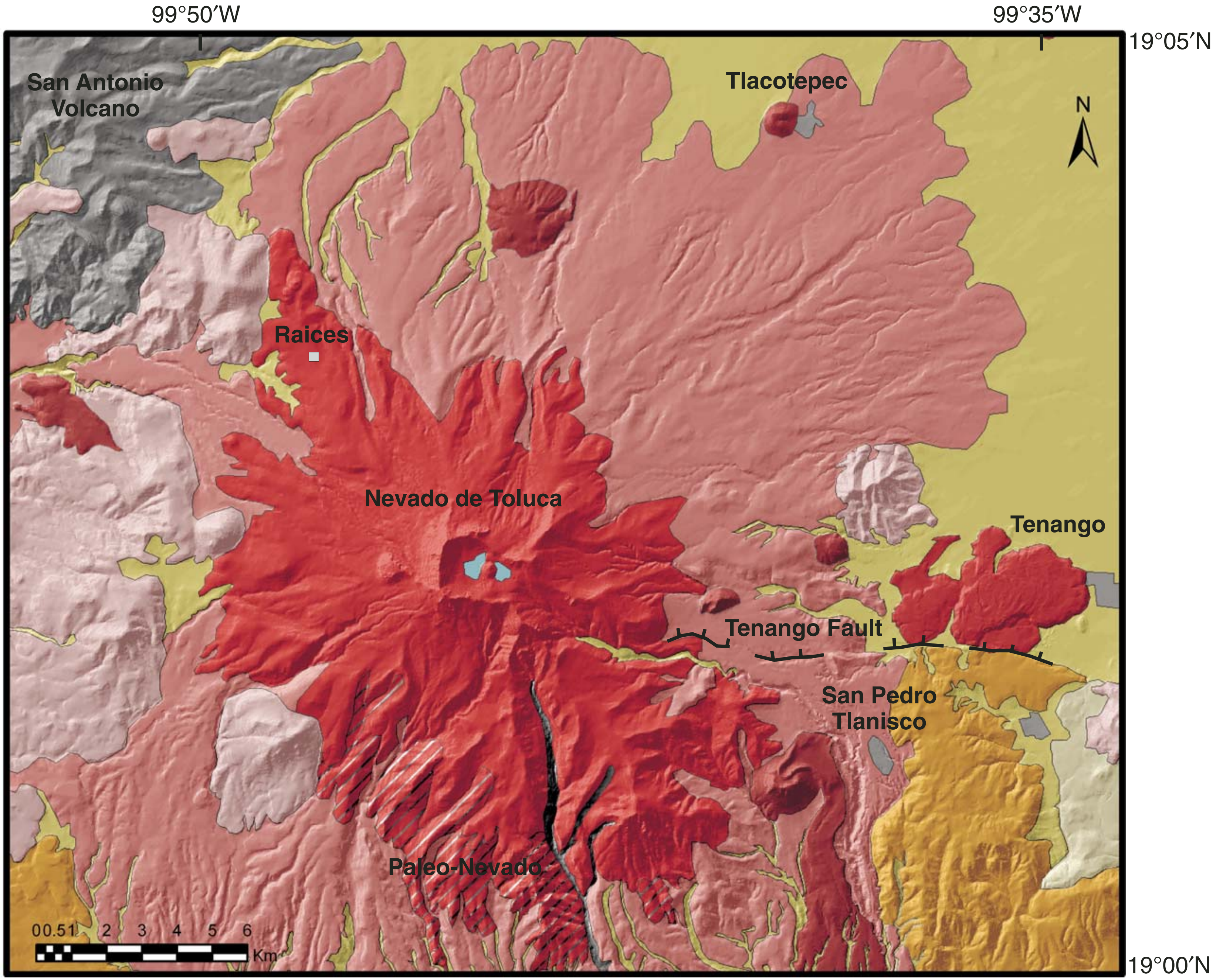


Figure 6 (*legend on following page*). Simplified geologic map of Nevado de Toluca volcano from García-Palomo et al. (2002) showing the basal sequence, San Antonio volcano, the units of Nevado de Toluca volcano, and the western portion of the Chichinautzin Volcanic Field.

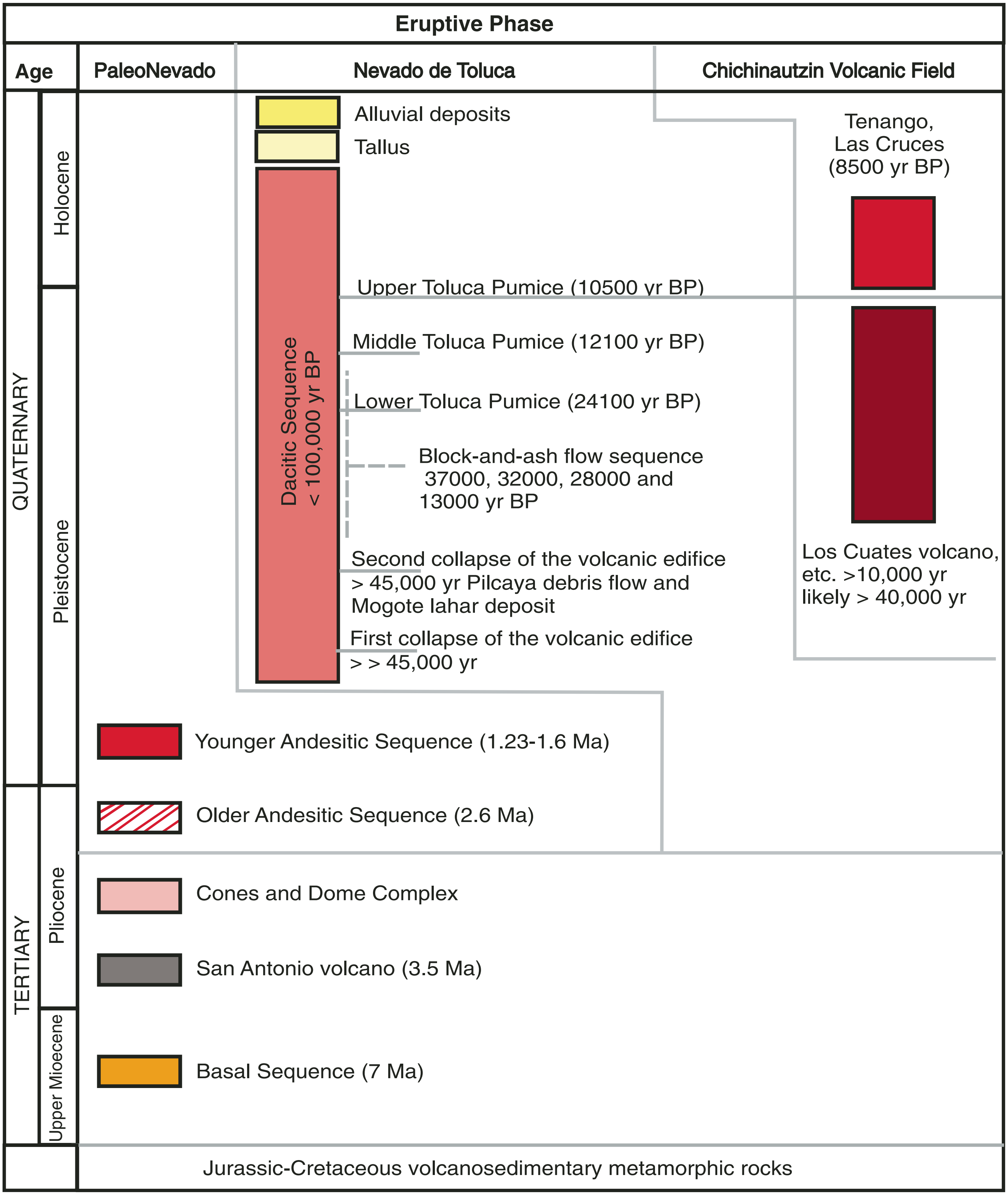


Figure 6 (*legend*).

Figure 7. (A) View of the western flank of Nevado de Toluca volcano showing the truncated crater with its maximum elevation represented by El Pico del Fraile. Photograph taken along the Toluca–Valle de Bravo highway by Jorge Neyra on 19 May 2001. (B) View to the southeast of the interior of Nevado de Toluca crater covered by snow in January 1992. Notice the dacitic El Ombligo dome and the frozen Lake of the Sun with its surface located at an elevation of 4230 m. In the background appears the volcano's crater open to the east.

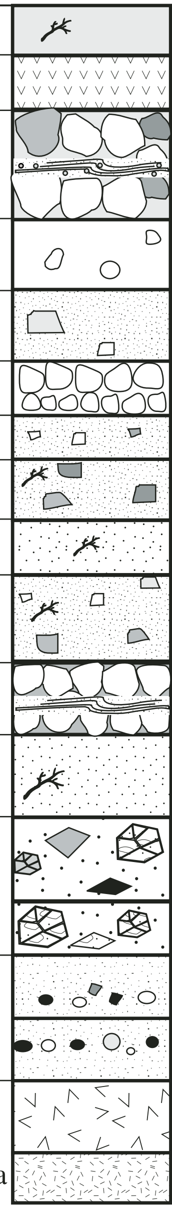
phology on the northern flanks and the infilling of graben type structures in the southern flank (García-Palomo, 1998; García-Palomo et al., 2002). The volcanic activity in the region began 1.6–1.3 Ma, with the formation of the ancient edifice called Paleonevado, which was located to the S-SE of the present edifice (Cantagrel et al., 1981), although recent studies suggest that the activity began 2.6 Ma (Norini et al., 2004). The Paleonevado edifice was constructed through the emplacement of andesitic lavas 1.2 Ma (García-Palomo et al., 2002; Norini et al., 2004). Between 1.2 Ma and 0.1 Ma, Paleonevado was subjected to intense erosive activity (Cantagrel et al., 1981) that generated debris avalanches, debris flows, and fluviatile deposits (Macías et al., 1997a; Capra and Macías, 2000) (Fig. 8). The formation of the present edifice of Nevado de Toluca started ca. 0.1 Ma, with the emission of dacitic products that have given place to explosive eruptions (Macías et al., 1997a; García-Palomo et al., 2002). During the late Pleistocene, the southern flank of Nevado de Toluca collapse twice, originating debris avalanche deposits that were transformed into debris flows with distance (Capra and Macías, 2000; Scott et al., 2001). The scars produced by these collapses have disappeared due to subsequent volcanic and glacier activity.

The oldest collapse produced a light brown, partially lithified debris avalanche deposit (described as DAD1 by Macías et al., 1997a). It has blocks with jigsaw-fit structures set in a coarse sand matrix with an approximate thickness of 10 m. This deposit has been found up to a distance of 35 km from the volcano summit (Fig. 5). The collapse was followed by the emplacement of lahars and fluviatile and lacustrine sediments, deposited in small, closed depressions over several thousands of years, although no specific dates are available. A second collapse of Nevado de Toluca also occurred during the late Pleistocene along with intense hydrothermal alteration of the volcanic edifice (Capra and Macías, 2000). This event generated a debris avalanche deposit, which was immediately transformed into a debris flow called "Pilcaya Debris Flow" that traveled 55 km from the source (Capra, 2000; Capra and Macías, 2002). The deposit is light-brown massive heterolithologic, with blocks immersed in a coarse sand matrix. This deposit rests upon a light brown paleosol and overlies a lahar deposit called "El Mogote," which was originated by the Pilcaya debris flow (Capra and Macías, 2002). According to their stratigraphic position, an age of >40,000 yr has been assigned to these deposits.

The activity of Nevado de Toluca during the late Pleistocene continued with two large explosive eruptions that produced two pink and white pumiceous pyroclastic flows, respectively, in all directions around the volcano. One of these deposits contains charcoal material dated at 42,000 yr. These deposits are generally exposed at the base of several gullies around the volcano. They are covered by deposits produced by the partial or total destruction of dacitic central domes and the generation of Plinian columns, as described in the following section.

Destruction of Dacitic Domes

Block-and-ash flows, produced by the partial or total destruction of domes, are recognized as "old lahar assemblages" by Bloomfield and Valastro (1974, 1977). These authors estimated that the age of this deposit is 28,000 yr B.P., as they dated a paleosol on top. However, Macías et al. (1997a) identified two block-and-ash flow deposits, the oldest one containing disseminated charcoal that yielded an age of 37,000 ± 1125 yr and therefore did not correlate with the Bloomfield and Valastro (1977) deposit. Macías et al. (1997a) correlated this deposit with the "gray lahar" deposit of Heine (1988), dated at 35,600 +2600/−800 yr, and a deposit overlying a paleosol, dated at 38,000 yr by Cantagrel et al. (1981). The younger block-and-ash flow deposit was dated at 28,140 +865/−780 and 28,925 +625/−580 yr B.P., with charcoal found inside the deposits that clearly correlated with the age of 27,580 ± 650 yr B.P. obtained by Bloomfield and Valastro (1974, 1977). Subsequent studies have described at least five gray massive block-and-ash flow deposits dated at ca. 37, 32, 28, 26, and



Thickness (meters)	Unit	Age (yr. B.P.)	Description
1.5		~3.3 ka	Gray cross-stratified surge and brown ash flow deposits with disseminated charcoal.
100		8.5 ka	Andesite lava flows (Tenango) of the Chichinautzin Formation.
20	**UTP**	10.5 ka	Upper Toluca Pumice. Fall deposit composed of three members, interbedded with thick pyroclastic flow and surge beds.
20	**MTP**	~12.1 ka	Sequence of pumice fall, surge and two white pumice flow deposits rich in subrounded dacitic pumice and biotite crystals.
5	**BAF**	~14 ka	Gray massive block-and-ash flow deposits, ash flows and surge layers with accretionary lapilli. The deposits contain juvenile dacitic clasts and pumice.
3	**LTP**	21.7 ka	Lower Toluca Pumice. Inversely graded fallout bed rich in yellow pumice and a few schist clasts from the local basement capped by surge beds.
10	**BAF**	~26.5 ka	Gray massive block-and-ash flow deposit. There are scarce outcrops of this deposit.
10	**BAF**	~28 ka	Gray massive block-and-ash flow deposit, composed of at least three units and interbedded surges. It contains juvenile dacitic clasts, pumice and red altered dacite clasts.
5	**BAF**	~32 ka	Pale brown ash flow deposit, composed of several flow units, interbedded with surge deposits.
10	**BAF**	~37 ka	Gray block-and-ash flow sequence composed of three main massive units and minor flow and surge layers. Consists of juvenile gray dacite clasts, red altered dacites, and rare pumice
3.5	Ochre Pumice Fall	~36-39 ka	Ochre pumice fall deposit, composed of three layers interbedded with surge deposits and caped by a massive pyroclastic flow rich in pinkish pumice fragments and charcoal.
~4		~42 ka	Pink pumice flow deposit, composed of several flow units. Clasts include subrounded dacitic pumice and few andesitic fragments, set in a sandy matrix.
40	**PDF**		Heterolithologic cohesive debris flow deposit (Pilcaya deposit), composed of dacitic clasts and exotic components (basalt, limestone, rhyolite, sandstone) embedded in an indurated sandy matrix
15	**DAD1**		Monolithologic debris avalanche deposit, composed of dacitic clasts set in a sandy matrix.
200	Older Sequence		Interbedded sequence of debris flows, hyperconcentrated flows, fluviatile beds, and minor lacustrine deposits "Older Lahars" from Nevado.
150		1.2-1.6 Ma	Primitive andesitic-dacitic lava flows of Nevado de Toluca.
100		2.6 ± 0.2 Ma	Light-gray porphyritic lava flows

Figure 8. Stratigraphic column of Nevado de Toluca after Macías et al. (1997a) and García-Palomo et al. (2002). The PDF (Pilcaya debris flow) and DAD1 (Debris Avalanche Deposit 1) are two debris avalanche-debris flow deposits produced by ancient collapses of the volcano. The young dacitic sequence of Nevado de Toluca rests on top of these deposits. BAF—block-and-ash flow; LTP—Lower Toluca Pumice; MTP—Middle Toluca Pumice; UTP—Upper Toluca Pumice.

14 ka (García-Palomo et al., 2002). The youngest (14 ka) deposit also appears in the stratigraphic record of the Lake Chiconahuapan (Newton and Metcalfe, 1999; Caballero et al., 2001, 2002). These five deposits have been widely distributed around Nevado de Toluca, and some of them (37, 28, and 14 ka B.P.) reach minimum distances of 25 km from the summit, still with considerable thicknesses (Fig. 9). The deposits contain juvenile dacitic lithics with silica concentrations varying from 65 to 67 wt%.

Plinian Eruptions

The first volcanological studies of Nevado de Toluca were dedicated to the deposits of two Plinian eruptions that were exposed on the flanks of the volcano and as far as the Lerma Basin. These deposits were dubbed by Bloomfield and Valastro (1974, 1977) as the Lower Toluca Pumice of ca. 24,500 yr B.P. and the Upper Toluca Pumice of ca. 11,600 yr B.P. These eruptions were dated with charcoal material below the deposits or the underlying paleosols, but never from material carbonized during the eruption. Recent studies suggest that these eruptions had younger ages, ca. 21,700 yr B.P. for the Lower Toluca Pumice (Capra et al., 2006) and ca. 10,500 yr B.P. for the Upper Toluca Pumice (Arce, 2003; Arce et al., 2003). However, the stratigraphy of Nevado de Toluca had some other deposits associated with Plinian- or subplinian-type activity. Macías et al. (1997a) discovered another deposit that yielded an age of ca. 12,100 yr B.P., dubbed the "White Pumice Flow," that was later described as the Middle Toluca Pumice by Arce et al. (2005b). In addition, García-Palomo et al. (2002) recognized another fallout dubbed as the Ochre Pumice dated between ca. 36,000 and 37,000 yr B.P. We next summarize the characteristics of the three youngest fall deposits of Nevado de Toluca:

The Lower Toluca Pumice (ca. 21,700 yr) was originated from a Plinian column that reached an elevation of 24 km above the crater and was later dispersed to the NE by stratospheric winds. The eruption was then followed by subplinian pulses and hydromagmatic eruptions that produced a total volume of 2.3 km^3 (0.8 km^3 dense rock equivalent [DRE]). This eruption has a peculiar place within the eruptive history of Nevado de Toluca, because it incorporated schist fragments from the basement, and the magma has a chemical composition that varies from andesitic 55wt% SiO_2 to dacite 65wt% SiO_2. The pumice of the Lower Toluca Pumice represents the more basic magma emitted by Nevado de Toluca during the past 50,000 yr (Capra et al., 2006).

Figure 9. El Refugio quarry located 15 km northeast of Nevado de Toluca crater showing an exposure of the 37,000 yr B.P. block-and-ash flow deposits.

The Middle Toluca Pumice (ca. 12,100 yr) began with a 20-km-high Plinian column, dispersed to the NE by stratospheric winds. The column was interrupted by hydromagmatic explosions and then an 18–19-km-high subplinian column established for some time prior to wane. Afterward, another subplinian column formed, and was interrupted by hydromagmatic explosions that produced pyroclastic surge, causing the column to collapse, triggering two pyroclastic flows rich in white pumice that were designated the "White Pumice Flow" by Macías et al. (1997a). This eruption generated 1.8 km^3 (DRE) of dacitic magma, with a homogeneous composition of 63.54–65.06 wt% in SiO_2. The mineral assemblage of this deposit is formed by phenocrysts of plagioclase > orthopyroxene > hornblende ± ilmenite and titanomagnetite and xenocrysts of biotite, immersed in a rhyolitic groundmass (70–71 wt% in SiO_2). The xenocrysts of biotite were in reaction with the groundmass; some of these were dated with the $^{40}Ar/^{39}Ar$ method and yielded ages of 0.8 Ma, indicating that they were assimilated from the magma chamber or other rocks (Arce et al., 2005a).

The Upper Toluca Pumice (ca. 10,500 yr) was caused by a complex event that formed four Plinian eruptions, PC0, PC1, PC2, and PC3, reaching heights of 25, 39, 42, and 28 km above the crater, respectively, that were dispersed toward the NE by dominant winds (Fig. 10). The last three columns were interrupted by hydromagmatic explosions at the crater that dispersed pyroclastic surges and produced the collapse of the columns, forming pumiceous pyroclastic flows. Fallouts PC1 and PC2 covered a minimum area of 2000 km^2, an area that is presently occupied by the cities of Toluca and México, and generated a volume of 14 km^3 (~6 km^3 DRE). In the basin of México, the Upper Toluca Pumice was initially described as the pumice triple layer (Mooser, 1967). The composition of the magma erupted by the Upper Toluca Pumice has a homogeneous composition of 63–66 wt% in silica (Arce et al., 2003).

Depth of the Magma Chamber

The homogeneous chemical composition and mineral assemblage of the magmas produced during the eruptions of 14,000 yr B.P. (block-and-ash flow), 12,100 yr B.P. (Middle Toluca Pumice), and 10,500 yr B.P. (Upper Toluca Pumice), suggest that the magma ejected by these eruptions might come from a single large reservoir filled with a dacitic magma (Arce et al., 2005b). With this assumption, these authors analyzed the mineral chemistry of the products of these three eruptions, which

Figure 10. Outcrop located 15 km northeast of Nevado de Toluca crater that reveals the 14,000 yr B.P. block-and-ash flow deposits. BAF14—Block-and-ash Flow 14; UTP—Upper Toluca Pumice.

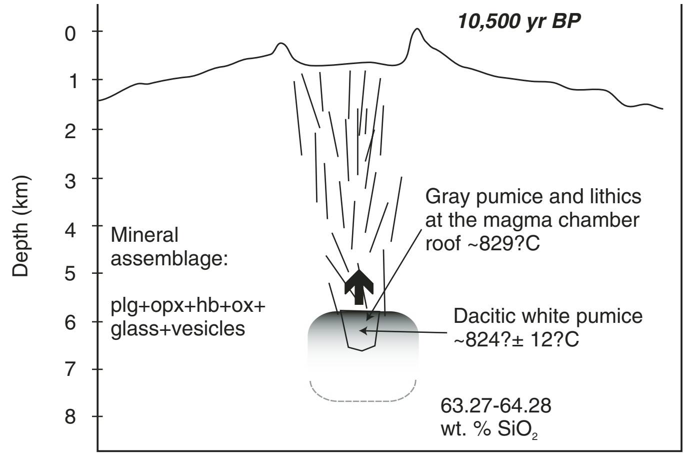


Figure 11. Cartoon showing the position of the magma chamber below Nevado de Toluca crater ca. 10,500 yr B.P. prior to the Plinian eruption that deposited the Upper Toluca Pumice (UTP). The UTP dacitic magma stagnated ~6 km below the crater at a temperature of 860 °C.

resulted to be very similar and consist of plagioclase ($An_{30–59}$) > orthopyroxene ($En_{56–59}$) > hornblende (edenita-hornblende mainly) >> Fe-Ti oxides + rare apatite (in orthopyroxene) + biotite, embedded in a rhyolitic groundmass (72–76 wt% in silica). For the ilmenite-titanomagnetite pair found in the samples, these authors estimated a pre-eruptive temperature of 850 °C and oxygen fugacity of −11. Hydrothermal experiments run on Upper Toluca Pumice samples found stability conditions for these magmas at pressures of 2 kbars at depths of ~6 km, beneath the Nevado de Toluca crater (Fig. 11).

The Holocene stratigraphic record of Nevado de Toluca consists of the extrusion of the Ombligo dacitic dome inside the crater. The dome surface has signs of glacier polish dated at 9100 ± 500 yr B.P., determined with the ^{36}Cl method (Arce et al., 2003). Therefore, it is clear that this dome was emplaced at the end of the Upper Toluca Pumice event or sometime later. The last eruption at Nevado de Toluca took place somewhere in the central crater ca. 3300 ^{14}C yr ago (Macías et al., 1997a). This eruption emitted a pyroclastic flow and surge on the NE flank of the volcano. Finally, a series of yellow lahar deposits are widely distributed around the volcano flanks. Nevado de Toluca should be considered to be an active dormant volcano.

POPOCATÉPETL VOLCANO

Popocatépetl volcano is located 65 km SE of México City and 45 km to the west of the city of Puebla (Figs. 1 and 12). More than one million people live in a radius of 40 km from the summit. The volcano forms the southern end of the Sierra Nevada range composed of the Tláloc, Telapón, Teyotl, Iztaccíhuatl, and Popocatépetl volcanoes. Popocatépetl means in náhuatl language “the smoking mountain,” referring to the fact that during prehispanic times, the Aztecs observed several eruptions such as those that occurred in 1363, 1509, 1512, and 1519–1528; the last one was also described by the priest Bernal Díaz y Gomarra (De la Cruz-Reyna et al., 1995). During the Colonial period, several descriptions were written about Popocatépetl eruptions, such as those that occurred in 1530, 1539, 1540, 1548, 1562–1570, 1571, 1592, 1642, 1663, 1664, 1665, 1697, and 1720. Throughout the nineteenth century, several scientists visited the volcano, among them von Humboldt in 1804 (Humboldt, 1862), the geologists Del Río in 1842, Del Castillo in 1870, and Sánchez in 1856, who carried out general descriptions of the volcano’s morphology. Aguilera and Ordóñez (1895) identified Popocatépetl as a truncated cone made of an alternating sequence of pyroclastic and lava deposits, with hypersthene-hornblende andesite composition. These authors pointed out the existence of seven fumaroles inside Popocatépetl’s crater with temperatures <100 °C and a blue-greenish lake formed by thaw water with variable temperatures of between 28 and 52 °C. Weitzberg (1922) made a detailed study of the glacier. In 1906, the bottom of the crater had a funnel-shaped with a small lake and vertical walls. Popocatépetl volcano reawakened in February 1919. In March-April, several inhabitants of Amecameca described that in the bottom of the crater there was an accumulation of steaming rocks, like a turned over stewing pan (Atl, 1939). On 11 October 1920, Waitz

Figure 12. Panoramic view of the southern flank of Popocatépetl volcano with snow covering up to the tree line at ~4000 m. The photo shows the surface of a debris avalanche deposit with hummocky topography.

visited the crater and described a lava dome at the bottom (Waitz, 1921). On 15 November 1921, Camacho and Friedlaender photographed the crater's interior, showing a lava dome surrounded by abundant gas emissions; they also captured the occurrence of brief explosions from the center of the dome (Friedlaender, 1921; Camacho, 1925). In January 1922, Camacho observed that a small crater was occupying the center of the dome. Atl (1939) described the eruption and evolution of this crater; according to his observations, the eruption ended in 1927. Popocatépetl remained calm for 67 yr; it reawakened on 21 December 1994.

Previous Studies

The few geological studies of Popocatépetl volcano prior to its reactivation in 1994 were those of Heine and Heide-Weise (1973), Miehlich (1984), Robin (1984), Carrasco-Núñez (1985), and Boudal and Robin (1989), as well as some petrologic studies (Boudal, 1985; Boudal and Robin 1987; Kolisnik, 1990). These studies defined Popocatépetl as a stratovolcano and presented part of the stratigraphic record. The geologic evolution of Popocatépetl can be summarized by the following stages: (1) the activity started with the formation of Nexpayantla volcano (Mooser et al., 1958), the ancestral volcano (Robin, 1984), through the emission of andesitic to dacitic lava flows. An eruption that occurred ~200,000 yr ago promoted the volcano collapse and the formation of a caldera. Inside this caldera began the construction of El Fraile volcano through the emplacement of andesitic and dacitic lava flows. This last volcano collapsed ca. <50,000 yr B.P. (Boudal and Robin, 1989), due to a Bezymiany-type eruption that destroyed the southern flank of the cone. These authors estimated a volume of 28 km^3 for this deposit. The eruption generated a debris avalanche that moved to the S-SW from the crater and was followed by the formation of a Plinian eruption that deposited a white pumice-fall layer toward the south of the volcano as well as pyroclastic flows. After this event, the formation of the modern cone known as Popocatépetl began. A large number of studies regarding geological (Siebe et al., 1995b, 1995c, 1996a, 1996b, 1997; Martín-Del Pozzo et al., 1997; Panfil et al., 1999; Espinasa-Pereña and Martín-del Pozzo, 2006) and geomorphological aspects (Palacios, 1996; Palacios et al., 2001) carried out during the last decade were driven by the reactivation of Popocatépetl. These stratigraphic studies can be summarized in the following sections.

Destruction of the Ancient Cone of Popocatépetl

Some 23,000 yr ago, a lateral eruption larger than the 18 May 1980 Mount St. Helens eruption produced the lateral collapse (to the south) of the ancient Popocatépetl cone (Fig. 13). The explosion generated a debris avalanche deposit that reached up to 70 km from the summit. The decompression of the magmatic system, due to the flank collapse, originated a lateral blast that emplaced a pyroclastic surge deposit and allowed the establishment of a Plinian column (Fig. 14). The column deposited a thick pumice-fall layer that is widely distributed on the southern flanks of the volcano. The column then collapsed and formed an ash flow that charred everything in its path. The deposit reached up to 70 km from the summit, covers an area of 900 km^2, and if we assign an average thickness of 15 m, a volume of 9 km^3 is obtained. This deposit overlies a paleosol that has charred logs dated at 23,445 ± 210 yr; disseminated charcoal found in the ash flow deposit yielded an age of 22,875 +915/−820 yr. Therefore the age of this eruption is ca. 23,000 yr B.P.

There are at least four debris avalanche deposits around Popocatépetl volcano. The oldest comes from the failure of the SE flank of Iztaccíhuatl volcano, and the other three come from the flank collapse of paleo-Popocatépetl (Siebe et al., 1995b; García-Tenorio, 2002), the youngest being the 23,000 yr B.P. deposit.

Construction of the Present Cone

Popocatépetl's present cone started growing 23,000 yr B.P.; it has a maximum elevation of 5472 masl and a relative elevation regarding the surrounding ground of 3000 m. The volcano has been built through the emission of alternating sequences of pyroclastic deposits and lava flows of andesitic-dacitic composition. The rocks consist of phenocrysts of plagioclase, hypersthene, augite, olivine, and rare hornblende in a glassy microcrystalline matrix. During the past 20,000 yr the explosive activity of Popocatépetl has been characterized by four major events (14,000, 5000, 2150, and 1100 yr B.P.) and four minor events (11,000, 9000, 7000 and 1800 yr B.P.) (Siebe et al., 1997; Siebe and Macías, 2006). The geologic history of Popocatépetl during the past 20,000 yr can be summarized in the following sections.

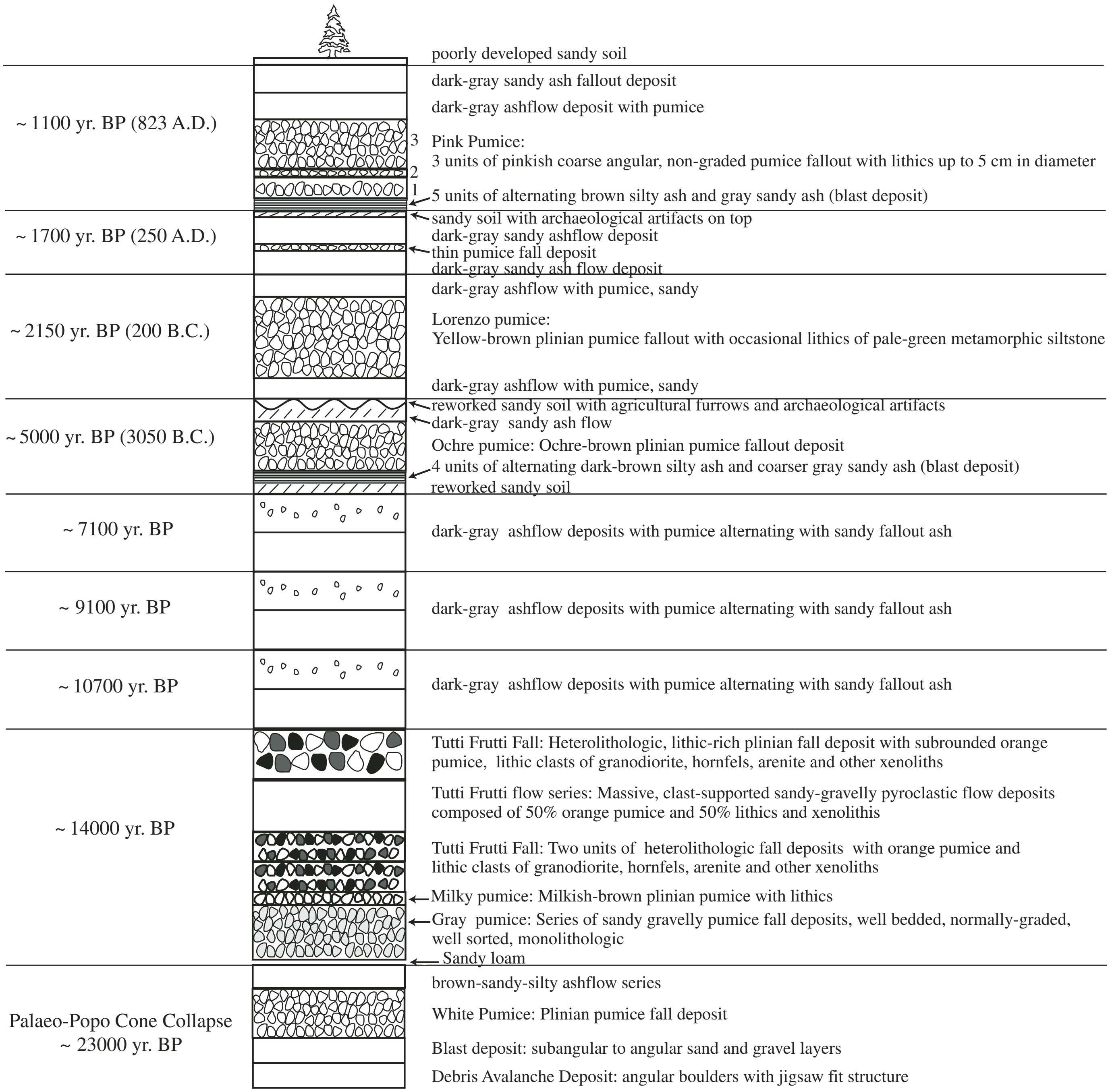


Figure 13. Simplified stratigraphic column of Popocatépetl volcano summarizing the pyroclastic deposits erupted during the past 23,000 yr B.P. (Siebe et al., 1995c; Siebe and Macías, 2006).

Phreatoplinian Eruption ca. 14,000 yr B.P.
(Pómez con Andesita or Tutti Fruti)

A large magnitude eruption began with the emission of a gray ash fallout around the volcano followed by a series of proximal pyroclastic flows and surges that culminated with the formation of a Plinian column. This was dispersed by the stratospheric wind to the N-NW, toward the present area occupied by México City (Siebe et al., 1995b, 1997). This fallout layer is heterolithologic; it contains ochre dacitic pumice, gray granodiorite, metamorphose limestones, skarns, and other fragments from the basement.

This fallout layer was described in the Basin of México as the "Pómez con andesita" or pumice with andesite by Mooser (1967), with a thickness of 5 cm. This deposit is widely exposed around the volcano, but it is barren of charcoal material except at

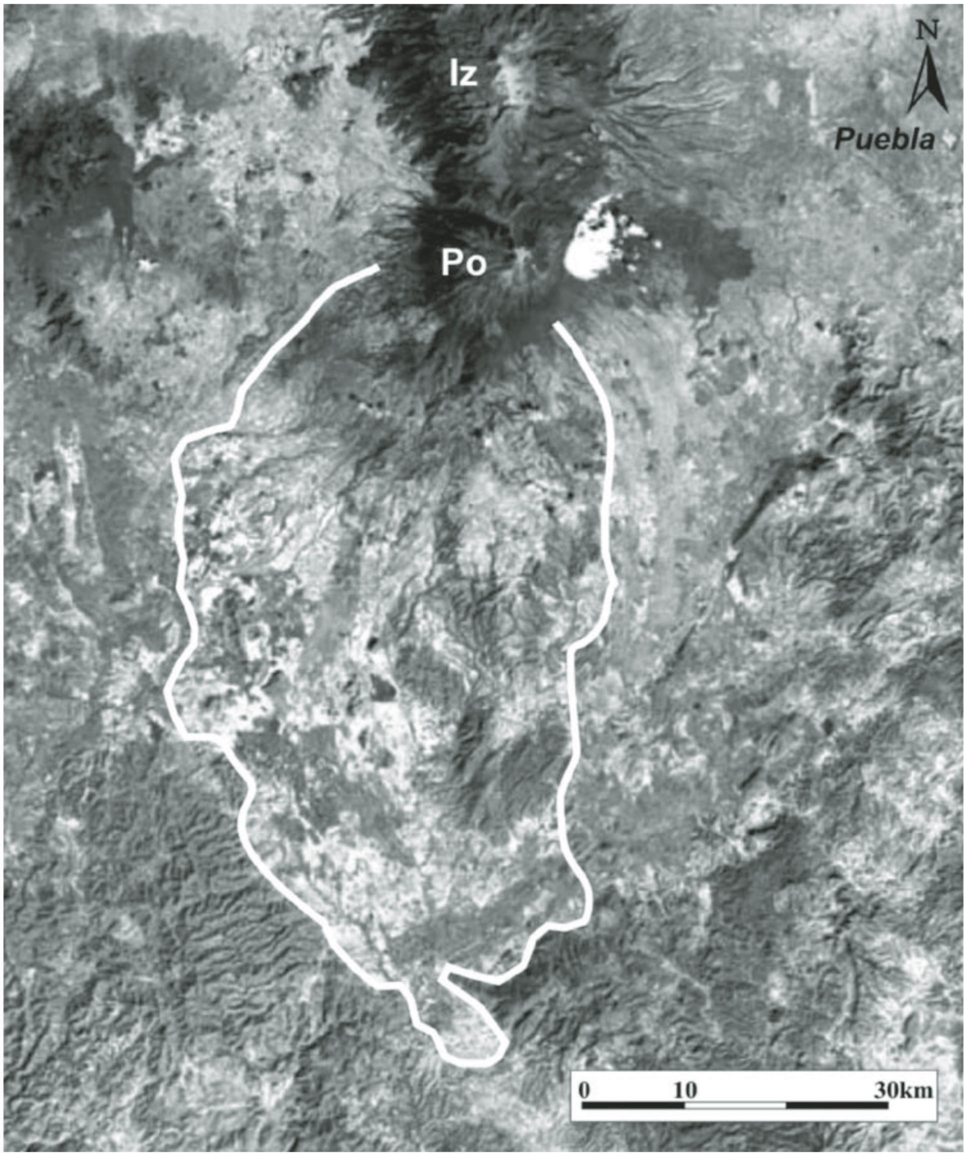


Figure 14. LANDSAT image that shows the Iztaccíhuatl (Iz) and Popocatépetl (Po) volcanoes and the distribution of the Popocatépetl debris avalanches according to Siebe et al. (1995b) (from Capra et al., 2002).

one site, where it yielded an age of ca. 14,000 yr B.P. (Siebe et al., 1997). This implies that at the time of the eruption Popocatépetl cone was covered by a glacier and therefore there was little vegetation on its flanks.

Recent Plinian Eruptions

At least three main Plinian fall deposits have been identified in the stratigraphic record of Popocatépetl volcano during the past 5000 yr. These deposits are dated at 5000, 2150, and 1100 yr B.P. (Fig. 15). The events that occurred at 5000 and 1100 yr B.P. had a similar evolution; they began with hydromagmatic explosions that dispersed wet pyroclastic surges up to 20 km from the summit. These explosions opened the magmatic conduit, decompressed the magmatic system, and formed >25-km-high Plinian columns (Siebe et al., 1996a, 1996b). These plumes were transported by the stratospheric winds to the N (5000 yr ago), to the E (2150 yr ago), and to the N-NE-E (1150 yr ago). Once these columns were fully established, their collapse was likely due to the consumption of the magma in the chamber, which generated pyroclastic flows that destroyed everything in their paths.

These eruptions blocked the hydrographic network of Popocatépetl and Iztaccíhuatl volcano located to the north. Springs and meteoric waters saturated the unconsolidated material produced by the pyroclastic flows to produce lahars that flooded the Basin of Puebla.

Figure 15. Photograph of the last three pumice fall deposits emplaced during the 5000, 2100, and 1100 yr B.P. eruptions at Popocatépetl volcano. C. Siebe points to the 2100 yr B.P. deposit.

Popocatépetl has also produced effusive activity through the emission of lava flows from the central event or from lateral fissures (Schaaf et al., 2005; Espinasa-Pereña and Martín-del Pozzo, 2006), such as those flows located in the vicinity of San Nicolás de Los Ranchos, which were produced during the eruption 2150 yr ago.

Effects of the Plinian Eruptions

The Popocatépetl Plinian eruptions had a large impact on prehispanic settlements through the direct emplacement of hot pyroclastic flows and thick fallouts or by the secondary emplacement of lahars that flooded extensive areas (Siebe et al., 1996a, 1996b). The last three Plinian eruptions of Popocatépetl coincide with three important events in Mesoamerican history: The 5000 yr B.P. (3195–2830 B.C.) eruption coincides with the 3114 B.C. beginning of the Mesoamerican Calendar; the 2150 yr B.P. (800–215 B.C.) eruption coincides with the transition from the Preclassic to the Classic period; the last eruption, which occurred 1100 yr B.P. (675–1095 A.C., likely 823 A.C.), coincides with the Classic-Postclassic periods transition (Fig. 16).

The 2150 yr B.P. Plinian eruption produced a yellow, pumice-rich fallout that was deposited on the NE flank of the volcano. The eruption emplaced a 1-m-thick fallout, burying prehispanic settlements in the region (Seele, 1973). Detailed archaeological excavations suggest that the inhabitants of this village, now called Tetimpa, were not able to get their belongings prior to their escape from the volcano's eruption (Plunket and Uruñuela, 1999, 2000, 2005).

The 1100 yr B.P. Plinian eruption also directly impacted prehispanic populations; however, the largest damage was produced due to remobilization of primary materials by lahars that moved along depth gullies. Thick sequences of lahars have been found surrounding Classic period ceremonial centers, such as Cholula, Cacaxtla, and Xochitécatl. The lahar deposits consist of the same constituents as the primary pyroclastic flow, surge-and-fall deposits exposed close to the volcano, and pottery shards, obsidian blades, and charcoal. The charcoal found in the lahar deposits yielded ages that correlate with the Plinian eruption at Popocatépetl volcano (675–1095 A.C.), with an approximated age of 823 A.C. Suárez-Cruz and Martínez-Arreaga (1993) concluded that the ceremonial center of Cholula was temporarily abandoned ca. 800 A.C.; that in a way coincides with the age range obtained by Siebe et al. (1996a, 1996b), therefore indicating that this abandonment of Cholula was caused by lahar flooding.

The 1994–2007 Eruption

After a quiescent period of 67 yr, Popocatépetl volcano resumed activity on 21 December 1994 with an increase in seismic activity and the emissions of 2–3-km-high columns above the crater, composed of ash, water vapor, and gases. The emissions were generated from small vents located at the base of the SE crater wall, inside the crater. These events continued sporadically until March 1995, when a sudden decrease in activity took place. From this time until early March 1996 the volcano remained calm with minor fluctuation in the seismic activity. On 29 March 1995, a lava flow at the base of the crater appeared for the first time. Small explosions launched lapilli-size fragments on the volcano flanks and in some villages. One of these explosions ended the

Figure 16. View of an archaeological excavation located on the northeastern slopes of Popocatépetl volcano. This small house forms part of the Tetimpa village covered by 1 m of lapilli-size pumice fall during the 2100 yr B.P. eruption of Popocatépetl.

lives of five mountaineers who had climbed the crater on 30 April 1996. This explosion sent juvenile material from the dome (1–2 cm in diameter) to the villages of Xalitzintla, San Nicolás de los Ranchos, and others located at 12 km on the NE flank of the volcano. On 10 June 1996, the dome had a thickness of 50 m and had completely covered the crater formed in 1922. Between April 1996 and June 1997, the volcano had extruded at least three lava domes. On 30 June 1997, a strong explosion preceded by two hours of volcanotectonic earthquakes took place. The eruption formed an 8-km-high column that was transported by the winds toward México City, producing ash fall that caused the international airport to be closed. The next day, a 12-km-long lahar was produced that reached the town of Santiago Xalitzintla, partially flooding one house. From July 1997 to November 2000, four domes were emplaced at the crater; these domes were followed by stronger explosions that launched ballistic projectiles up to 5 km from the summit. From 12–16 December 2000, Popocatépetl activity increased notably, causing the evacuation of more than 40,000 inhabitants. At that time, Popocatépetl's crater was almost completely filled with lava, and the 18 December 2000 eruption had extensive coverage in the media. These vulcanian-type explosions projected dome rocks >5 km from the summit, triggering forest fires. After these events, an updated hazards map of the volcano was presented (Sheridan et al., 2001). From April 1996 to the present, the volcano has extruded more than 20 domes that were subsequently destroyed by vulcanian-type explosions. One of the last large explosions took place on 21 January 2001, producing a small column that suddenly collapsed to produce a pyroclastic flow with a maximum extent of 6.5 km, reaching to the edge of the tree line. The pyroclastic flow diverted into several tongues, one of which scoured the glacier forming, and lahars reached the town of Xalitzintla, located 15 km from the summit (Capra et al., 2004). After 11 yr of activity, Popocatépetl's crater is almost filled to the brim, thus overcoming the magnitude of the 1919–1927 eruption (Macías and Siebe, 2005) (Fig. 17).

PICO DE ORIZABA VOLCANO (CITLALTÉPETL)

Pico de Orizaba volcano represents the highest mountain in México (19°01′N, 97°16′W, 5675 m) (Fig. 18A). This volcano is also known as Citlaltépetl, which means in náhuatl language "Mountain of the Star," and is located in the eastern portion of the Trans-Mexican Volcanic Belt (Fig. 1). Its summit sets the boundary between the states of Veracruz and Puebla. There are several reports of the historic activity of Pico de Orizaba; the most important event occurred in 1687 (Mooser, et al., 1958), and reports of other small events have been summarized by De la Cruz-Reyna and Carrasco-Núñez (2002). Today, there are a few signs of minor activity at the volcano, such as weak emissions of SO_2 and sulfur sublimates in the inner crater walls (Waitz, 1910–1911). For this reason, it is considered an active but dormant volcano. The present edifice has a central oval crater with 500 × 400 m diameter and 300 m vertical walls (Fig. 18B). The crater's northern part is covered by a glacier (Heine, 1988). The cone has a symmetric shape with steep slopes that reach up to 40° (Fig. 18A). The volcano rises 2900 m with respect to the Serdán–Oriental Basin to the west and 4300 m with regard to the Coastal Gulf Plain to the east (Carrasco-Núñez, 2000).

Figure 17. Detailed view to the east of Popocatépetl crater's interior. The photograph was taken on 24 October 2004 by Jorge Neyra. At the center appears a 340-m-wide internal crater. In the background to the right is Pico de Orizaba volcano.

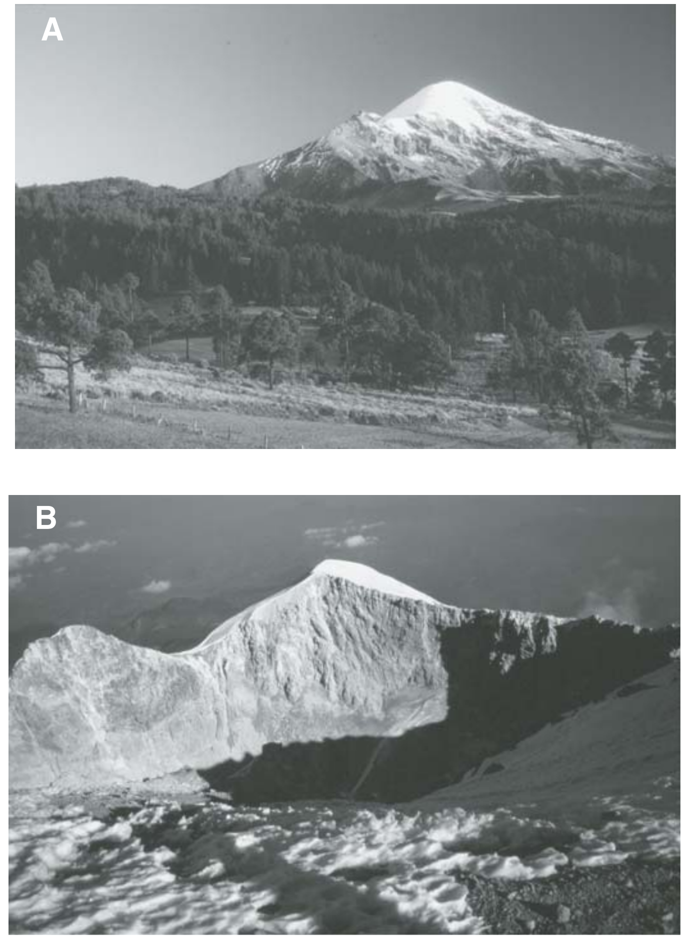


Figure 18. (A) Photograph of the western flank of Pico de Orizaba volcano seen from Tlalchichuca on 4 February 2004. Photograph by Jorge Neyra. (B) View from the summit of the E-NE internal wall of Pico de Orizaba crater. Photograph by Jorge Neyra taken on 8 November 1998.

Previous Studies

The first general observations of the volcano were made by Waitz (1910–1911) during a geologic field trip; after that, the volcano remained unstudied until the middle of the twentieth century, when it was considered an active volcano by Mooser et al. (1958). The first regional geological studies of the volcano were carried out by Yáñez-García and García-Durán (1982) and Negendank et al. (1985). The first stratigraphy of Pico with a general evolution model was presented by Robin and Cantagrel (1982). This work was followed by petrologic studies (Kudo et al., 1985; Singer and Kudo, 1986; Calvin et al., 1989), stratigraphic and eruptive chronology studies (Cantagrel et al., 1984; Hoskuldsson, 1992; Hoskuldsson et al., 1990; Hoskuldsson and Robin, 1993; Carrasco-Núñez, 1993, 1997; Carrasco-Núñez et al., 1993; Siebe et al., 1993; Carrasco-Núñez and Rose, 1995; Gómez-Tuena and Carrasco-Núñez, 1999; Rossotti and Carrasco-Núñez, 2004), and the completion of a geological map (Carrasco-Núñez and Ban, 1994; Carrasco-Núñez, 2000). All of this information allowed scientists to produce a volcano hazards map (Sheridan et al., 2002, 2004), applying different programs, such as FLOW3D (Kover, 1995) and LAHARZ (Iverson et al., 1998). Finally, Díaz-Castellón (2003) and Zimbelman et al. (2004) evaluated the stability of the volcanic edifice.

Eruptive History

Pico de Orizaba is a Quaternary volcano built upon Cretaceous limestones and claystones (Yáñez-García and García-Duran, 1982). Its eruptive history shows several episodes of construction and destruction of ancient cones. In fact, Robin and Cantagrel (1982) and Hoskuldsson (1992) proposed that the volcano was built during three main stages. However, subsequent stratigraphic studies indicated that Pico de Orizaba was built during four eruptive phases (Carrasco-Núñez and Ban, 1994; Carrasco-Núñez, 2000) (Fig. 19). These phases were dubbed from older to younger as the Torrecillas cone, the Espolón de Oro cone, silicic peripheral domes, and the Citlaltépetl cone, as described in the following:

1. The Torrecillas cone began its formation 0.65 ± 0.71 Ma (Hoskuldsson, 1992) with the emission of the Pilancon olivine basaltic andesites, followed by the Jamapa andesitic and dacitic lavas and the complex Torrecillas two pyroxene andesitic lavas, breccias, amphibole-bearing dacites, dated at 0.29 ± 0.5 Ma (Carrasco-Núñez, 2000), with two pyroxene andesites atop. The Torrecillas stratovolcano reached a total volume of 270 km^3 (Carrasco-Núñez, 2000).

The constructive stage of this volcano ended with the collapse of the edifice, which generated a debris avalanche (Hoskuldsson et al., 1990) called Jamapa (Carrasco-Núñez and Gómez-Tuena, 1997) ca. 0.25 Ma. The debris avalanche traveled 75 km to the east of the crater, along the Jamapa River (Carrasco-Núñez and Gómez-Tuena, 1997) (Fig. 20). This phase produced a caldera, whose remnants are exposed to the south of the present cone (Robin and Cantagrel, 1982; Carrasco-Núñez, 1993).

2. The Espolón de Oro cone was constructed to the north of the Torrecillas crater walls; the remnants of this structure are represented by amphibole-dacite dated at 0.21 ± 0.04 Ma (Carrasco-Núñez, 2000). The Espolón de Oro cone began its formation with the emission of plagioclase + amphibole andesitic lavas dubbed Paso de Buey; these were followed by the Espolon de Oro amphibole dacitic lavas and block-and-ash flows. The activity of the volcano continued first on the western flank with the lateral emission of El Carnero olivine basaltic andesites and on the northern flank with the emplacement of the Alpinahua sequence composed of pyroclasts and andesitic lavas between 0.15 and 0.09 Ma. This sequence ended with the emplacement of aphanitic andesitic lavas interbedded with breccias and a welded ignimbrite with fiamme structures. At this point, the Espolón de Oro cone had an approximated volume of 50 km^3 (Carrasco-Núñez, 1997). A flank collapse finished with the construction of the Espolón de Oro cone ca. 16,500 yr B.P. (Carrasco-Núñez et al., 2005). This collapse originated the Tetelzingo debris avalanche deposit; that downslope transformed into a cohesive debris flow (10%–16% clay) that traveled 85 km from the source, covered and area of 143 km^2, and had a volume of 1.8 km^3 (Carrasco-Núñez et al., 1993). These authors concluded that the collapse was due to hydrothermal alteration of the volcanic edifice and the presence of a summit glacier.

3. Silicic peripheryal domes: The domes Tecomate to the NE and Colorado to the SW (Fig. 20) were extruded during the construction of the Espolón de Oro cone. The Tecomate dome consists of rhyolitic obsidian lavas, while the Colorado dome is formed by dacitic lavas and associated pyroclastic flows. These domes were followed by the emplacement of the Sillatepec and Chichihuale dacitic domes to the NW of the cone. This stage ended with the emplacement of the Chichimeco dome complex, composed of domes and amphibole andesitic lavas covered by a scoria pyroclastic flow dated at 8630 ± 90 yr B.P. (Carrasco-Núñez, 1993).

4. The construction of the Citlaltépetl cone started ca. 16,500 yr B.P. within the Espolón de Oro caldera wall remnants. The cone began its construction with effusion of the hornblende dacite Malacara lavas that flowed 13 km to the SE flank. These lavas were followed by the emission toward the NE flank of the crater of the Vaquería, andesitic lavas, and finally by the emission of the thick Orizaba dacites toward the SW and NE flanks. The present cone has a volume of 25 km^3 (Carrasco-Núñez, 1997). The historic lavas that erupted in 1537, 1545, 1566, and 1613 (Crausaz, 1994) are well exposed around the summit (Carrasco-Núñez, 1997).

The formation of the Citlaltépetl cone is not solely related to the quiet effusion of andesitic-dacitic lavas on its flanks, but it is also related to explosive activity (Siebe et al., 1993; Hoskuldsson and Robin, 1993; Carrasco-Núñez and Rose, 1995; Rossotti and Carrasco-Núñez, 2004). The stratigraphic record shows at least three major explosive events: (1) an eruption produced pyroclastic pumice flows on the eastern flanks of the volcano ca. 13,000 yr B.P.; (2) the Citlaltépetl eruptive sequence that produced diverse pumice fallouts and pyroclastic flows between 8500 and 9000 yr B.P. (Carrasco-Núñez and Rose, 1995); and (3) a series of dome destruction events that emplaced block-and-ash flow deposits on the western and southeastern flanks of the volcano occurred ca. 4100 yr B.P. (Siebe et al., 1993; Carrasco-Núñez, 1999). Other pyroclastic flow deposits have been dated between 8170 and 1730 yr B.P., and six pyroclastic fall deposits have been dated between 10,600 and 690 yr B.P. (De-la Cruz-Reyna and Carrasco-Núñez, 2002). Among all the deposits described above, only the 8500–9000 and 4100 yr B.P. eruptions left traceable deposits around the crater. Because of their importance in the stratigraphic record of Pico de Orizaba, these are described in the following sections.

Citlaltépetl Ignimbrite (8500–9000 yr B.P.)

This sequence represents the most explosive event during the Holocene record of Pico de Orizaba. The sequence was originally described as two members composed of pyroclastic flows separated by pumice fallout (Carrasco-Núñez and Rose, 1995). These authors concluded that the lower member was formed by four pyroclastic flow units with charcoal dated at 8795 ± 57 yr B.P. (average of six dates), a lahar, and a poorly developed paleosol. The upper member was composed of a fall deposit at the base and a pyroclastic flow deposit with charcoal dated at 8573 ± 79 yr B.P. (average of ten dates). Carrasco-Núñez and Rose (1995) proposed that these eruptions occurred between 8500 and 9000 yr B.P. at the Citlaltépetl cone. The pyroclastic flows (Citlaltépetl Ignimbrite) produced during this event were dispersed around the crater up to a distance of 30 km. The deposits consist of andesitic scoria, minor pumice, and lithics set in a fine ash matrix and have a volume of 0.26 km^3 (Carrasco-Núñez and Rose, 1995). These authors concluded that the Citlaltépetl Ignimbrite was triggered by a magma mixing event between andesitic (58–59 wt% SiO_2) and (62–63 wt% SiO_2) dacitic magmas and emplaced through a boiling-over mechanism. Afterward, Gómez-Tuena and Carrasco-Núñez (1999) studied the pyroclastic flow deposit found in the lower member of the Citlaltépetl Ignimbrite. According to its distribution, granulometry, and textural features, these authors concluded that the ignimbrite settled through an accretion process in proximal facies and in masse in distal facies. Recent studies of the Holocene deposits of Pico de Orizaba have recognized at least 10 fallout layers, interbedded with four pyroclastic flows and three paleosols (Rossotti and Carrasco-Núñez, 2004; Rossotti, 2005). Carrasco-Núñez and Rose (1995) dated three samples of organic material in this sequence that gave ages falling in the 8500–9000 yr B.P. period, the same period proposed for the Citlaltépetl Ignimbrite. Based on new stratigraphic columns, correlation with previous work, and radiometric dating, Rossotti (2005) concluded that the Citlaltépetl sequence occurred in four eruptive phases, dated at ca. 9000–8900, ca. 8900–8800, ca. 8800–8700, and ca. 8700–8500 yr B.P. This means that in a 500 yr period, Citlaltépetl volcano produced four eruptions that emplaced fallouts to the NE and pyroclastic flows up to a distance of 30 km from the crater.

Dome Destruction Event (4100 yr B.P.)

This event was characterized by the destruction of a central dome that originated block-and-ash flow deposits that formed a pyroclastic fan in the outskirts of the Avalos village west to the crater (Siebe et al., 1993). The flows traveled ~16 km and had an H/L (elevation loss/lateral extent ratio) of 0.186. Siebe et al. (1993) estimated a volume of 0.048 km^3 for these deposits and obtained radiocarbon dates of 4040 ± 80 and 4060 ± 120 yr B.P. In 1999, Carrasco-Núñez described a sequence of block-and-ash flow deposits forming terraces to the southeast of the crater. These flows traveled ~28 km from the crater and have an H/L of 0.153. Carrasco-Núñez (1999) determined a volume of 0.162 km^3 and obtained a radiocarbon date of 4130 ± 70 yr B.P. Because both deposits have similar ages and a homogeneous chemical composition of the juvenile lithics (dacite 62.7–63.95 wt% SiO_2), Carrasco-Núñez (1999) proposed that these deposits, with a total volume of 0.27 km^3, were emplaced by the same eruption ca. 4100 yr B.P.

EL CHICHÓN VOLCANO

El Chichón volcano (17°21′N, W93°41′W; 1100 masl) is located in the NW portion of the State of Chiapas, at ~20 km to the SW of the Pichucalco village. El Chichón is the active volcano in the Chiapanecan Volcanic Arc (Damon and Montesinos, 1978). This arc straddles the State of Chiapas from the southeast at the Boquerón volcano to the northwest at El Chichón volcano. The Chiapanecan Volcanic Arc is composed of low-relief, small volume volcanoes of Pliocene to Recent age (Capaul, 1987) and is located between the Trans-Mexican Volcanic Belt and the Central America Volcanic Arc (Fig. 1). From a tectonic point of view, the volcano sits on the northern part of the strike-slip Motagua-Polochic fault Province (Meneses-Rocha, 2001). El Chichón is built upon Late Jurassic–Early Cretaceous evaporites and limestones, Early to Middle Cretaceous dolomites, and Tertiary claystones and limestones (Canul and Rocha, 1981; Canul et al., 1983; Duffield et al., 1984; García-Palomo et al., 2004) (Fig. 21). These rocks are folded in a NW-SE direction, forming the Catedral anticline and La Unión and Caimba synclines (Macías et al., 1997b; García-Palomo et al., 2004). In addition, these folded rocks are dissected by E-W left lateral faults, such as the San Juan Fault that cuts across the volcano and N45°E-trending Chapultenango normal faults, dipping to the NW (García-Palomo et al., 2004).

El Chichón consists of the 1.5 × 2 km wide Somma crater, which has a maximum elevation of 1150 masl (Fig. 22). This crater is a dome ring structure breached in its N, E, and SW flanks that has subvertical inner walls and gentler outer slopes. The relative elevation of the Somma crater with respect to the surrounding terrain is ~700 m to the east and 900 m to the west. The Somma crater is composed of trachyandesites with an age of 0.2 Ma (Damon and Montesinos, 1978; Duffield et al., 1984).

The Somma crater was disrupted by younger structures, such as the Guayabal tuff cone in the SE, and intruded by two trachyandesitic domes of unknown age exposed in the SW and NW flanks (Macías, 1994). Prior to the 1982 eruption, this Somma crater was occupied by a trachyandesitic dome with two peaks, the higher having an elevation of 1235 masl. The 1982 eruption destroyed this dome, forming a 1-km-wide crater with a maximum elevation of 1100 masl (Fig. 23). This crater has vertical walls up to 140 m deep, and its floor is at an elevation of 860 m, hosting a lake, fumaroles, mud, and boiling water ponds (Taran et al., 1998; Tassi et al., 2003; Rouwet et al., 2004). The water in the lake normally has a temperature of 32 °C (Armienta et al., 2000), and the fumaroles have temperatures of 100 °C (Taran et al., 1998; Tassi et al., 2003), with organic components (Capaccioni et al., 2004) connected to an active hydrothermal system (Rouwet et al., 2004) (Fig. 24).

Previous Studies

The first mention of El Chichón volcano was in 1930, when local inhabitants heard and felt noises on a small hill known as Cerro la Union or Chichonal (Müllerried, 1933). Müllerried visited the hill and described that, in fact, it was an active volcano composed of a crater (Somma crater) and a central dome. In addition, he found a set of fumaroles with temperatures ~90 °C and a small lake between the SE part of the dome and the crater (Somma crater), thus showing that it was an active volcano. El Chichón appeared later in the catalogue of active volcanoes of the world (Mooser et al., 1958). For several decades, the volcano remained unstudied until, in the 1970s, the National Power Company (Comisión Federal de Electricidad, CFE) began a geothermal prospection study of the area (González-Salazar, 1973; Molina-Berbeyer, 1974). During a mineral prospecting study of the State of Chiapas, Damon and Montesinos (1978) visited the volcano; they dated the eastern wall of the Somma crater at 0.209 ± 0.019 Ma (K-Ar method) and considered it to be an active

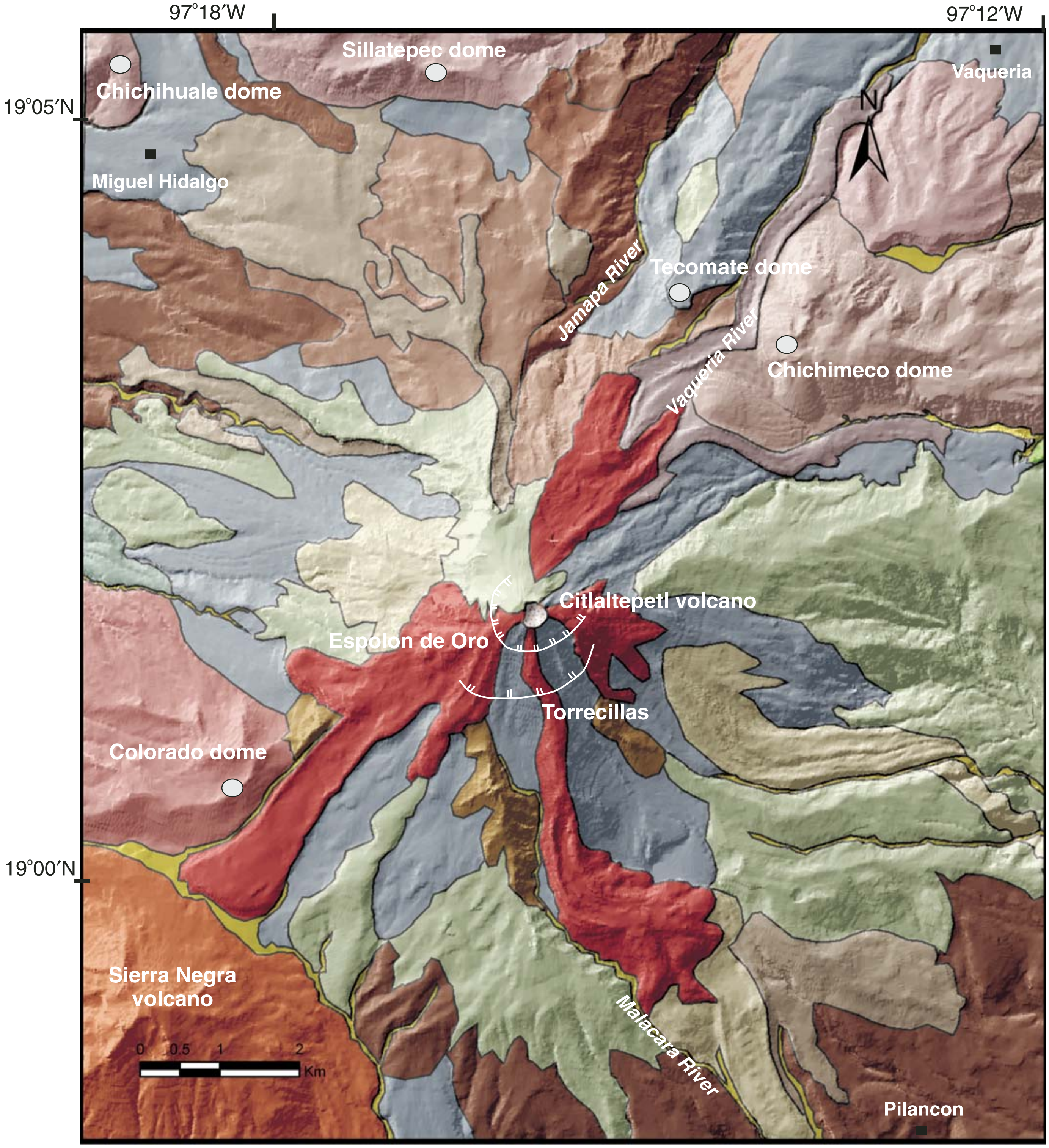


Figure 19 (*legend on following page*). Simplified geological map of Pico de Orizaba volcano showing its four stages of evolution (from Carrasco-Núñez, 2000).

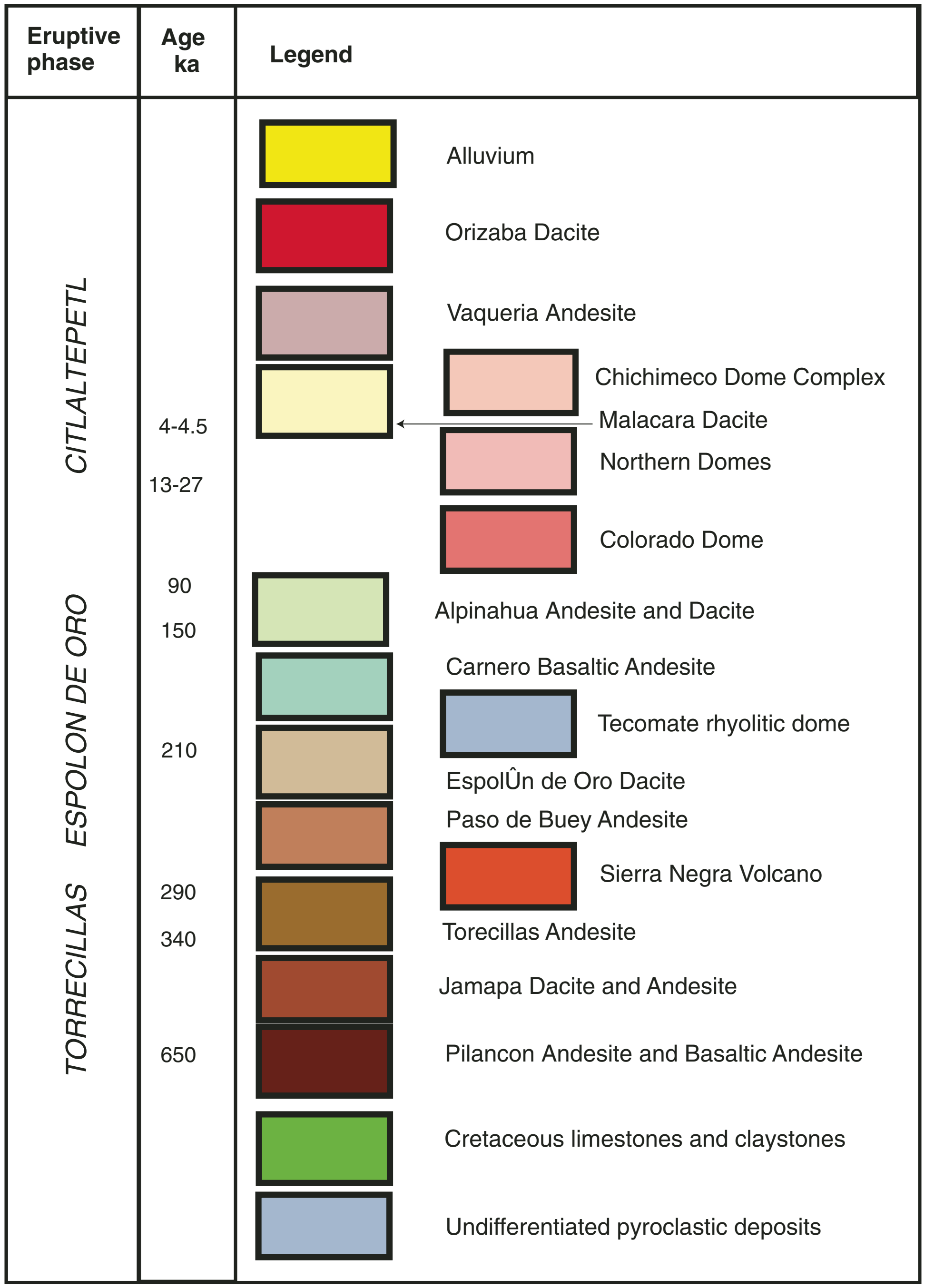


Figure 19 (*legend*).

volcano. At the beginning of the 1980s, the CFE started a geological reconnaissance (Canul and Rocha, 1981) and a fluid study (Templos, 1981) of the area. In their geological study, Canul and Rocha (1981) recognized older deposits rich in charcoal related to previous eruptions. While pursuing fieldwork in 1981, they felt and heard strong explosions at the volcano and realized that El Chichón might reactivate in the near future, as they stated in their internal report to the CFE.

The 1982 Eruption

Prior to the 1982 eruption, El Chichón consisted of the Somma crater and a central dome known as Chichón or Chichonal (Fig. 22). The eruption of El Chichón came as a surprise to the local population and the distant scientific community in México City, despite the increasing premonitory signs given by the volcano since late 1981.

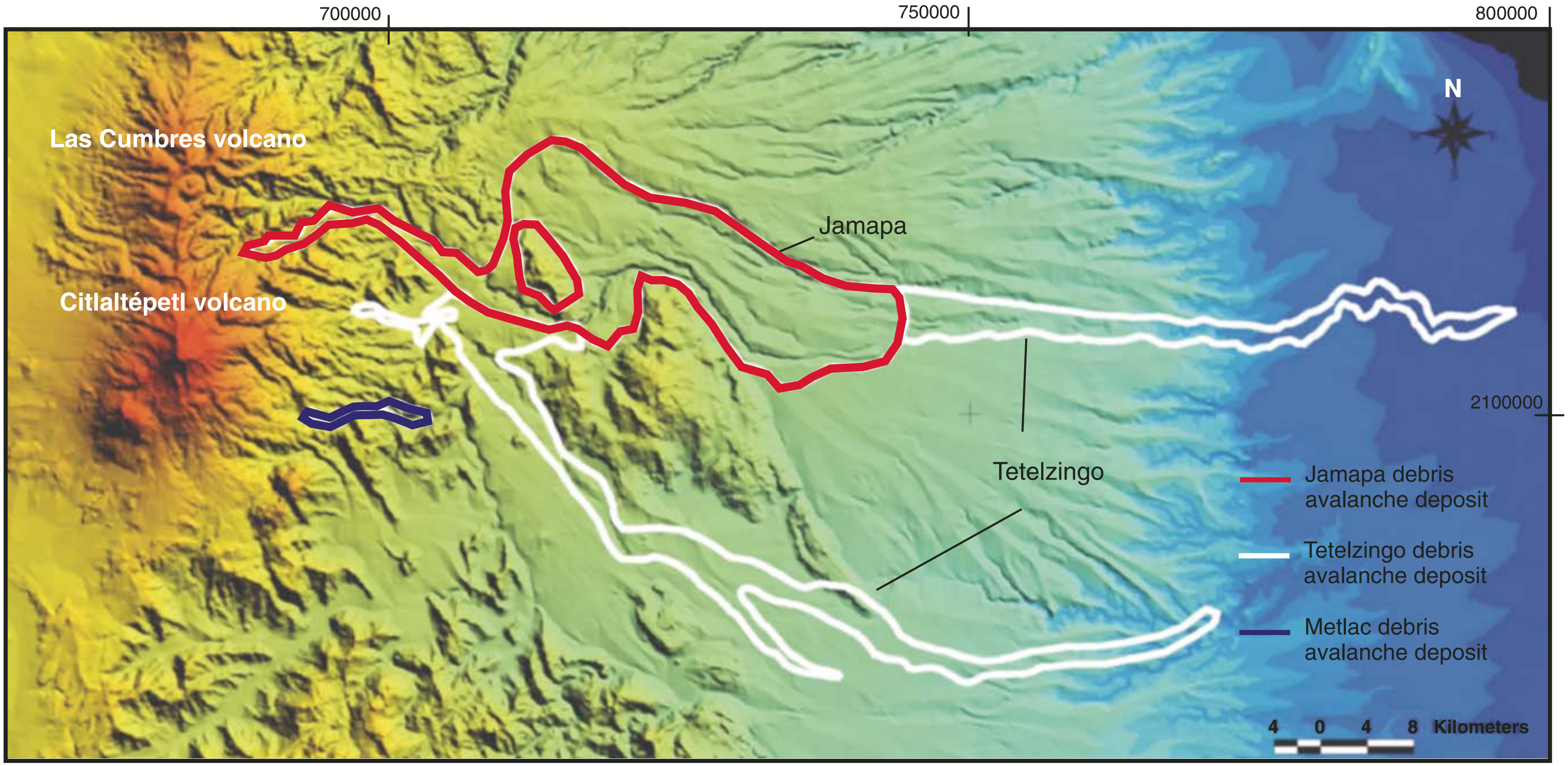


Figure 20. Digital elevation model showing the distribution of debris avalanches and debris flows produced at Pico de Orizaba volcano (Carrasco-Núñez et al., 2006). Topographic database with UTM coordinates obtained with Geosensing Engineering and Mapping.

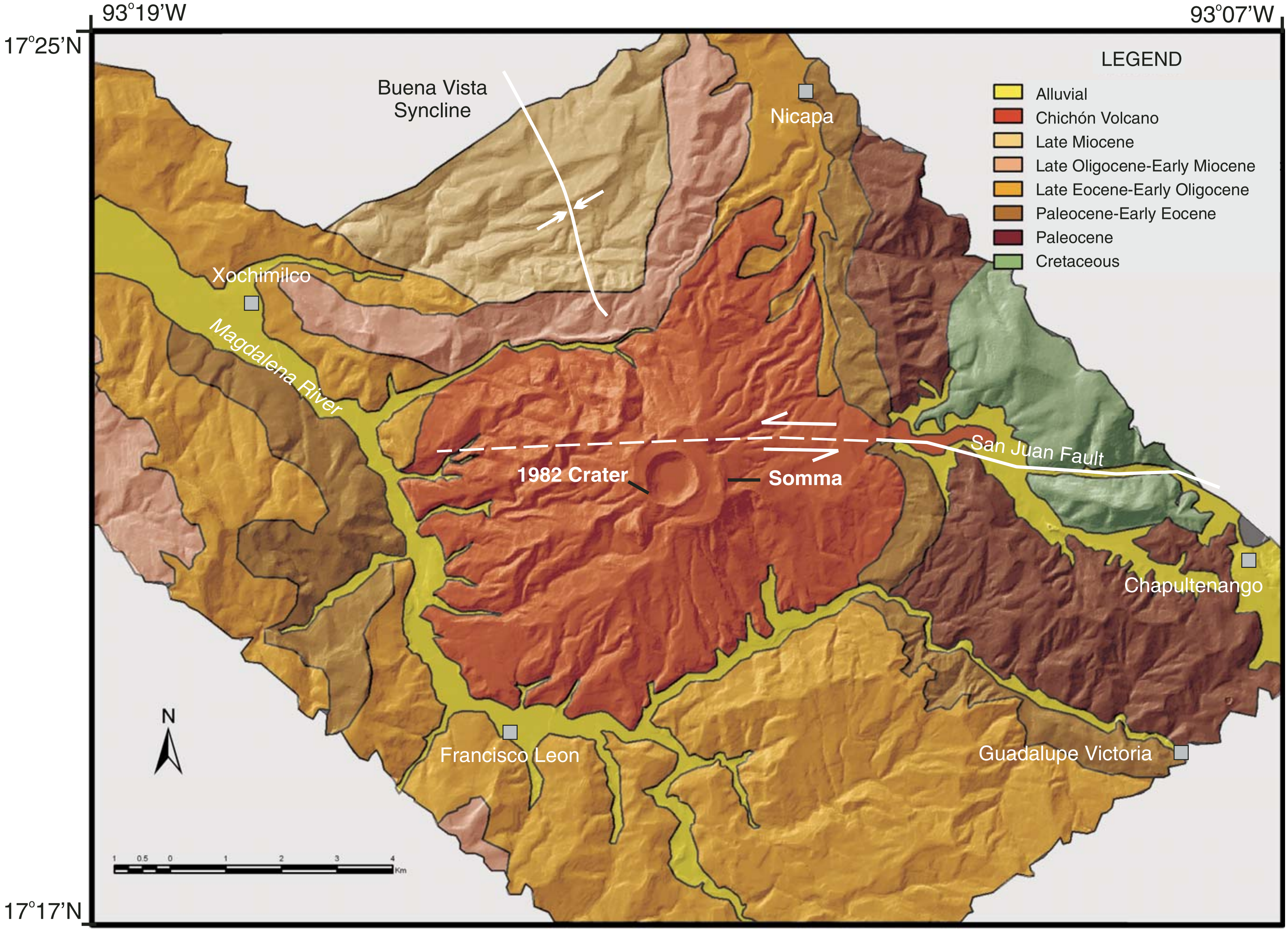


Figure 21. Simplified geological map of El Chichón volcano with the presence of the Buena Vista syncline and the left-lateral San Juan Fault (taken from García-Palomo et al., 2004).

Figure 22. View to the west of El Chichón volcano with the Somma crater (arrow) and the central dome prior to the 1982 eruption. Photograph courtesy of Rene Canul.

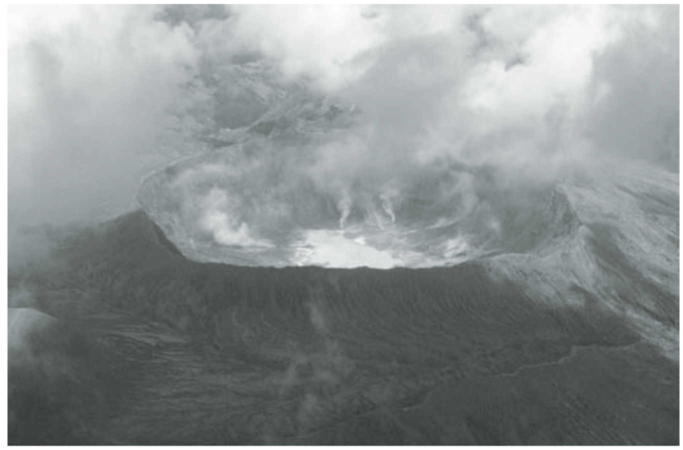


Figure 23. Aerial view of El Chichón volcano after the 1982 eruption showing the crater that was formed in 1982 and an internal lake with fumaroles and sulfur precipitation. Photograph courtesy of Reynaldo Mota.

Figure 24. Panoramic view from the SE of the El Chicón crater lake taken in May 2004. Fumarolic activity is observed on the righthand side of the picture.

The eruption was preceded by fumarolic activity and earthquakes; the latter were recorded at a seismic network installed at the Chicoasén hydroelectric plant, under construction at that time by the CFE (Espíndola et al., 2002). Jiménez et al. (1999) concluded that the seismic activity started in late 1980, increased through 1981, and peaked in March 1982. On Sunday night, 28 March 1982, the seismic activity developed into harmonic tremor of changing amplitude, followed by an hour of complete calm. Then, the beginning of the eruption was recorded as large amplitude tremor. The magmatic eruption opened a 150–180-m-wide crater in the central dome and formed a 27-km-high Plinian column (Medina-Martínez, 1982; Sigurdsson et al., 1984; Carey and Sigurdsson, 1986). This column was dispersed to the NE by the stratospheric winds, depositing fall layer A. The ash fall provoked confusion amongst the inhabitants, and they fled toward the nearest towns and cities of Pichucalco and Tuxtla Gutiérrez in Chiapas State and Villahermosa in Tabasco State. The next day, the area was closed by the army, which immediately applied its emergency plan called DN-III-E (Secretaría de la Defensa Nacional, 1983), during which most of the villagers were evacuated. From 29 March to 2 April, the volcano's activity decreased despite the occurrence of minor explosions and constant seismic activity. On Saturday, 3 April 1982, the military authorities allowed villagers to return to their homes, mainly to the Francisco Leon village, located in the southwestern part of the volcano, where a geologist and an army convoy spent the night. Unfortunately, the more violent explosion occurred that night (Yokoyama et al., 1992). The trachyandesitic magma came into contact with groundwater and the hydrothermal system, producing a series of hydromagmatic eruptions that dispersed pyroclastic surges (S1) up to 8 km from the crater, destroying nine towns and killing ~2000 people (Sigurdsson et al., 1984; Carey and Sigurdsson, 1986; Sigurdsson et al., 1987; Macías et al., 1997c; Macías et al., 1998; Scolamacchia and Macías, 2005) (Fig. 25). These explosions completely destroyed the central dome, first allowing the formation of a lithic-rich pyroclastic flow (F1) that partially flattened the rugged terrain, then the establishment of a 32 km high phreatoplinian column depositing layer fall B, rich in lithics, the collapse of the column with the generation of pyroclastic flows and surges rich in pumice (F2 and S2), and finally a series of flow and surge events restricted to 2 km from the crater (UI). The volcano stayed relatively calm for four hours, until dawn on 4 April, when a 29-km-high Plinian column developed. The plume was dispersed to the NE by the stratospheric winds, which deposited layer C. After this, hydromagmatic explosions took place and emplaced pyroclastic surges (S3) up to 4 km from the crater. The eruptive plumes contained water vapor, as proved by the aggregation of ash (Varekamp et al., 1984) and the formation of different types of ash aggregates (Scolamacchia et al., 2005). The eruption left a 1-km-wide crater with four small inner craters with lakes. The walls of the crater exposed a stratigraphic record that clearly suggested this crater had been active in the past. After 4 April, El Chichón activity decreased drastically to small explosions, until September 1982.

The April 1982 eruption emitted 1.5 km^3 of magma (DRE) and abruptly modified the hydrologic network around the volcano. The pyroclastic deposits blocked all gullies around the volcano, especially the Susnubac-Magdalena and Platanar riverbeds (Riva-Palacio Chiang, 1983; Secretaría de la Defensa Nacional, 1982). Pyroclastic flows (F1 and F2) produced on the night of 3 April formed a 25–75-m-thick dam at the Magdalena River (Macías et al., 2004a). The water coming from the Susnubac River and rainfall during the months of April and May accumulated behind the dam, forming a lake (Fig. 26). The pyroclastic material in the lake substrate had a temperature of ca. 300 °C, and the lake water almost reached the boiling point. At the end of April, the lake was 4 km long, 300–400 m wide, and had a volume of hot water of 26 × 106 m^3, and in early May it had achieved a volume of 40×10^6 m^3 (Medina-Martínez, 1982). In the last days of May, the army evacuated 1288 inhabitants living downstream from the dam (Báez-Jorge et al., 1985) because the water from the lake started overtopping the dam. On 26 May 1982 at 1:30 a. m., the dam broke, producing two subsequent debris flows that rapidly were transformed into a single hyperconcentrated flow (Macías et al., 2004a). At 10 km from the dam, the flow had temperature of 82 °C, and reached the town of Ostuacán, burning coffee and cacao plantations. Downstream, the flow became a laden sediment flow, and at the junction between the Magdalena and Grijalva rivers, it diluted, promoting the rise of the water column. After 7 km, the flow front reached the wall of the Peñitas hydroelectric plant of CFE, at that time under construction. At the Peñitas dam, the surface of the river rose 7 m, with a surge of hot water (50 °C) that killed one worker, injured three others, and destroyed machinery.

The 1982 eruption devastated more than 100 km^2 of tropical forest and in distant places, such as Ostuacán and Pichucalco, caused churches and house roofs to collapse. The eruption injected seven million tons of SO_2 into the atmosphere, forming aerosols (Krueger, 1983; Matson, 1984). The aerosols formed a cloud that circumnavigated the planet and dropped the global temperature by ~0.5 °C. The juvenile products of the magma contained anhydrite, a mineral that had never been reported as primary mineral in magmas (Luhr et al., 1984; Rye et al., 1984). All these ingredients attracted the attention of a large number of specialists, who studied the crater lake (Casadevall et al., 1984), the pyroclastic deposits (Tilling et al., 1984; Sigurdsson et al., 1984, 1987; Rose et al., 1984), the ash and plume dispersion (Varekamp et al., 1984; Carey and Sigurdsson, 1986), the sulfur content of the magma (Devine et al., 1984; Carroll and Rutherford, 1987), and the petrology of the products (Luhr et al., 1984).

Holocene Eruptions

The 1982 eruption stripped all vegetation cover around the volcano and exposed the stratigraphic record of El Chichón. The first stratigraphic studies of this volcano revealed pyroclastic deposits with charcoal produced by eruptions that occurred at 550, 1250, and 1650 yr B.P. (Rose et al., 1984; Tilling et al., 1984). Surprisingly, Tilling et al. (1984) found abundant Mayan-type pottery shards embedded in the 1250 yr B.P. deposits, suggesting that the volcano had been inhabited by Mayan groups since that time. Later, Macías (1994) found the deposits of other two eruptions occurring at the volcano, dated at 900 and 1400 yr B.P. In the first systematic stratigraphic study of El Chichón, Espíndola et al. (2000) discovered at least 11 deposits related to the same number of eruptions during the past 8000 yr at 550, 900, 1250, 1400, 1700, 1800, 2000, 2400, 3100, 3700, and 7500 yr B.P. (Fig. 27). The 550, 1250, and 1450 yr B.P. eruptions were larger in magnitude than the 1982 eruption, having a volcanic explosivity index (VEI) of at least 4 (Newhall and Self, 1982). In fact, the 550 yr B.P. Plinian eruption ejected 1.4 km^3 of magma through the emplacement of a single fall layer that is thicker if compared to the total fallouts A, B, and C, produced by the 1982 eruption at the same distance (Macías et al., 2003). Strikingly, the 2400 yr B.P. pyroclastic flow also contained pottery shards, suggesting that the volcano was inhabited during the past 2500 yr. Unfortunately, these shards are homemade artifacts, and therefore it was not possible to assign them an Olmec or Maya origin, two of the oldest cultures in the region. The El Chichón eruptions not only had a local impact but also a regional one through the emplacement of falling ash at the Maya Low Lands in Guatemala during the Classic period (Ford and Rose, 1995). The inhabitants of this region used volcanic ash as a temper to prepare pottery. The chemical composition of the crystals and glass found in the pottery shards is similar to the content of juvenile materials located at El Chichón and Tacaná volcanoes in México and other volcanoes in Guatemala such as Cerro Quemado (Ford and Rose, 1995).

The repose period between the Holocene eruptions of El Chichón varies from 100 to 600 yr, compared with the 1982 eruption, which occurred after 550 yr of quiescence (Tilling

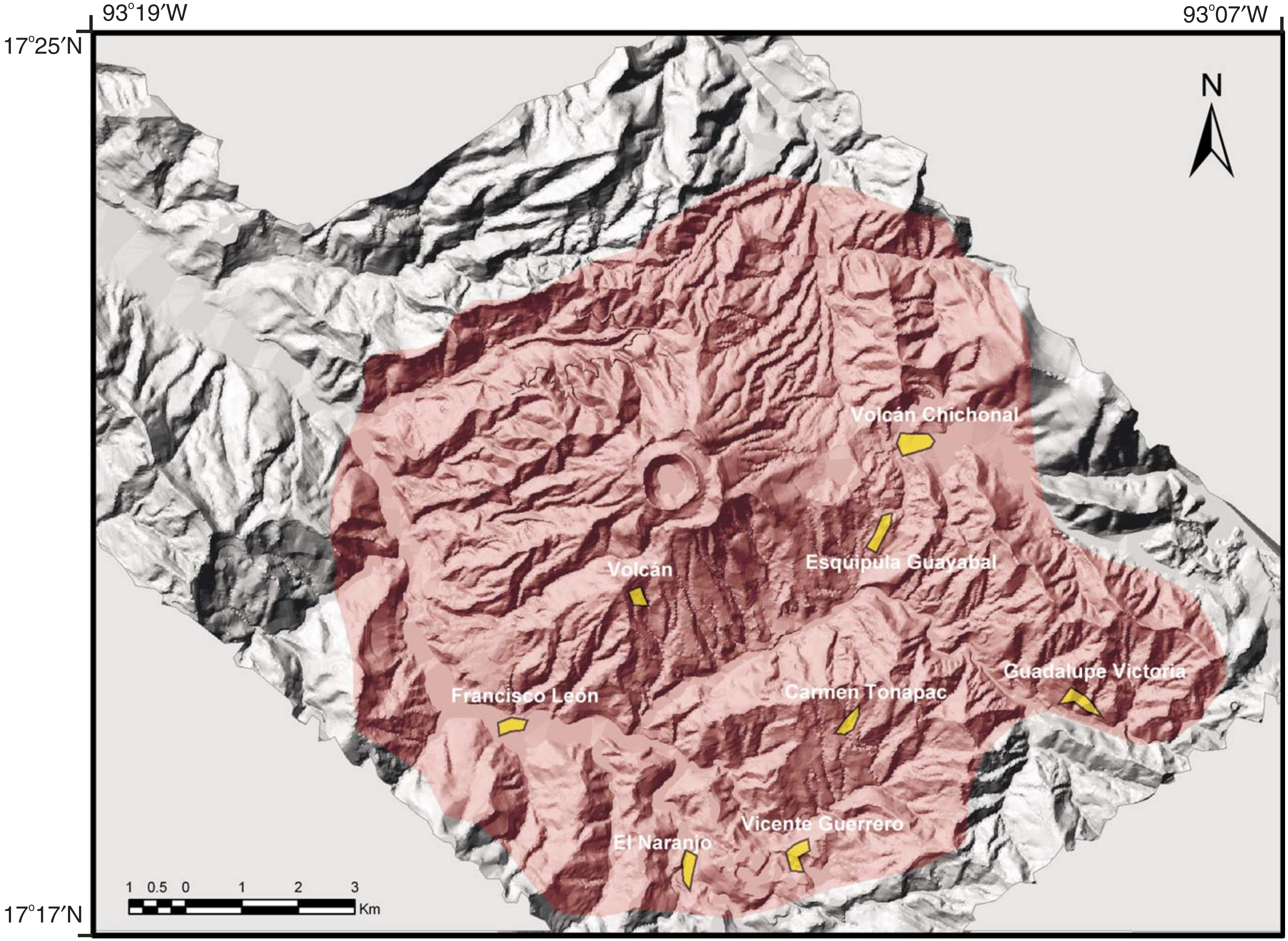


Figure 25. Digital elevation model of El Chichón volcano displaying the distribution of the pyroclastic surges around the volcano (after Scolamacchia and Macías, 2005). The towns in yellow were destroyed during the 1982 eruption.

et al., 1984; Espíndola et al., 2000). Therefore, today we cannot discount that the volcano might erupt under the present circumstances with an open crater occupied by a lake. In fact, the 900, 2000 and 2400 yr B.P. eruptions occurred under open vent conditions.

Chemical Composition of the Magmas

The magmas erupted by El Chichón during the past 8000 yr have a trachyandesitic composition (Duffield et al., 1984; Rose et al., 1984; McGee et al., 1987; Espíndola et al., 2000; Macías et al., 2003), with a similar mineral association of plagioclase > amphibole > augite, with magnetite, sphene, pyrrothite, biotite, and apatite as accessory minerals (Luhr et al., 1984; Duffield et al., 1984; Espíndola et al., 2000). The chemical composition and the mineral association are very similar to those described for the Chiapanecan Volcanic Arc rocks (Capaul, 1987). The 1982 trachyandesites were rich in sulfur and crystals (~53 vol.%) including ~2 vol.% of anhydrite (Luhr et al., 1984). Prior to the eruption, this magma had temperatures between 750 and 880 °C (Luhr et al., 1984) and stagnated at depths ~6 km (2 kilobars, Luhr, 1990). However, the analysis of the seismic record indicates that the magma chamber at the time of the eruption was at a depth between 7 and 13 km below the volcano (Jiménez et al., 1999), suggesting a deeper reservoir in the sedimentary Jurassic-Cretaceous substrate where assimilation of sulfur from the rocks could have taken place (Rye et al., 1984). Prior to the 550 yr B.P. eruption, the magma had a temperature of 820–830 °C, it was water saturated (5–6 wt% H_2O), and it resided at a depth of ~6–7.5 km (2–2.5 kilobars) below the volcano (Macías et al., 2003).

The homogeneous chemical composition of the Holocene Chichón products, suggests that the magmatic system has been relatively stable. However, the presence of mafic enclaves (trachybasalts and trachybasaltic andesites) hosted by the trachyandesites of the Somma crater and the isotopic variations found in the plagioclase phenocrysts of El Chichón show that the volcano magmatic system has had the input of repeated recharges of deeper mafic magmas (Espíndola et al., 2000; Tepley et al., 2000; Davidson et al., 2001; Macías et al., 2003) (Fig. 28).

TACANÁ VOLCANO

Tacaná volcano (15°08′N, 92°09′W) lies in the State of Chiapas in southern México and in the San Marcos Department in Guatemala. Shared by the two countries, the summit delineates the international boundary. The volcano is located at the most northern part of the Central American Volcanic Arc (Figs. 1 and 29). For several small towns and coffee plantations, this volcano is a serious hazard to their population and economic activity. The city of Tapachula, Chiapas, México (~300,000 inhabitants) is located 30 km SW of the volcano's summit.

Previous Studies

Von Humboldt (1862) first described the volcano as Soconusco, the most northwestern volcano of Central America, while Dollfus and de Monserrat (1867) referred to it as Istak volcano. Later, Sapper (1896, 1899) clarified that Soconusco was synonym of Tacaná, the name used by Böse during his studies (1902, 1903, 1905). Finally, Waitz (1915) con-

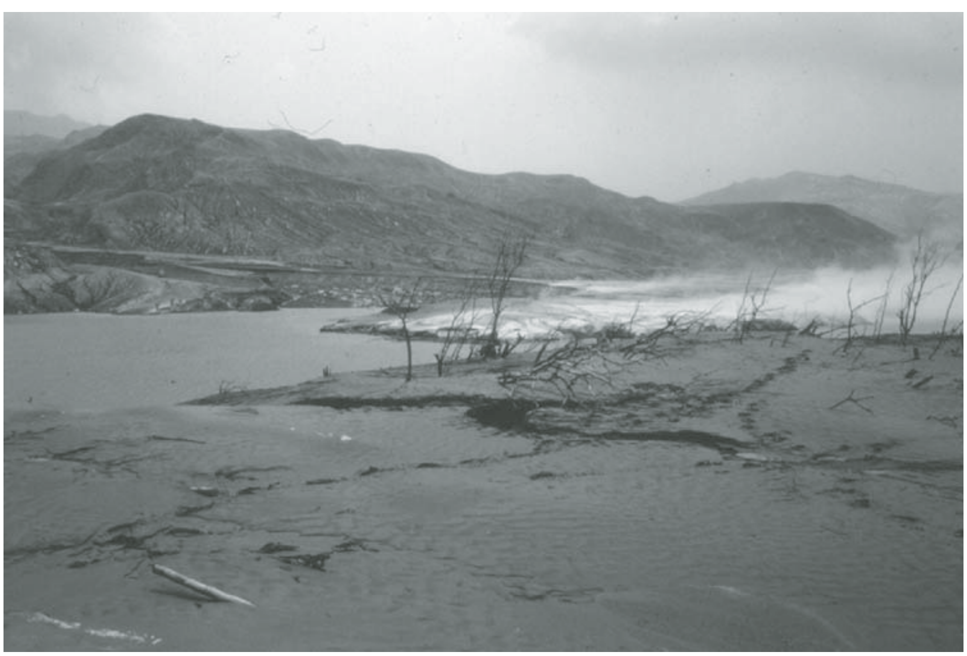


Figure 26. View of the temporal lake formed in April to May 1982. The water in contact with the pyroclastic deposits was nearly to the boiling point. Photograph courtesy of Servando de la Cruz.

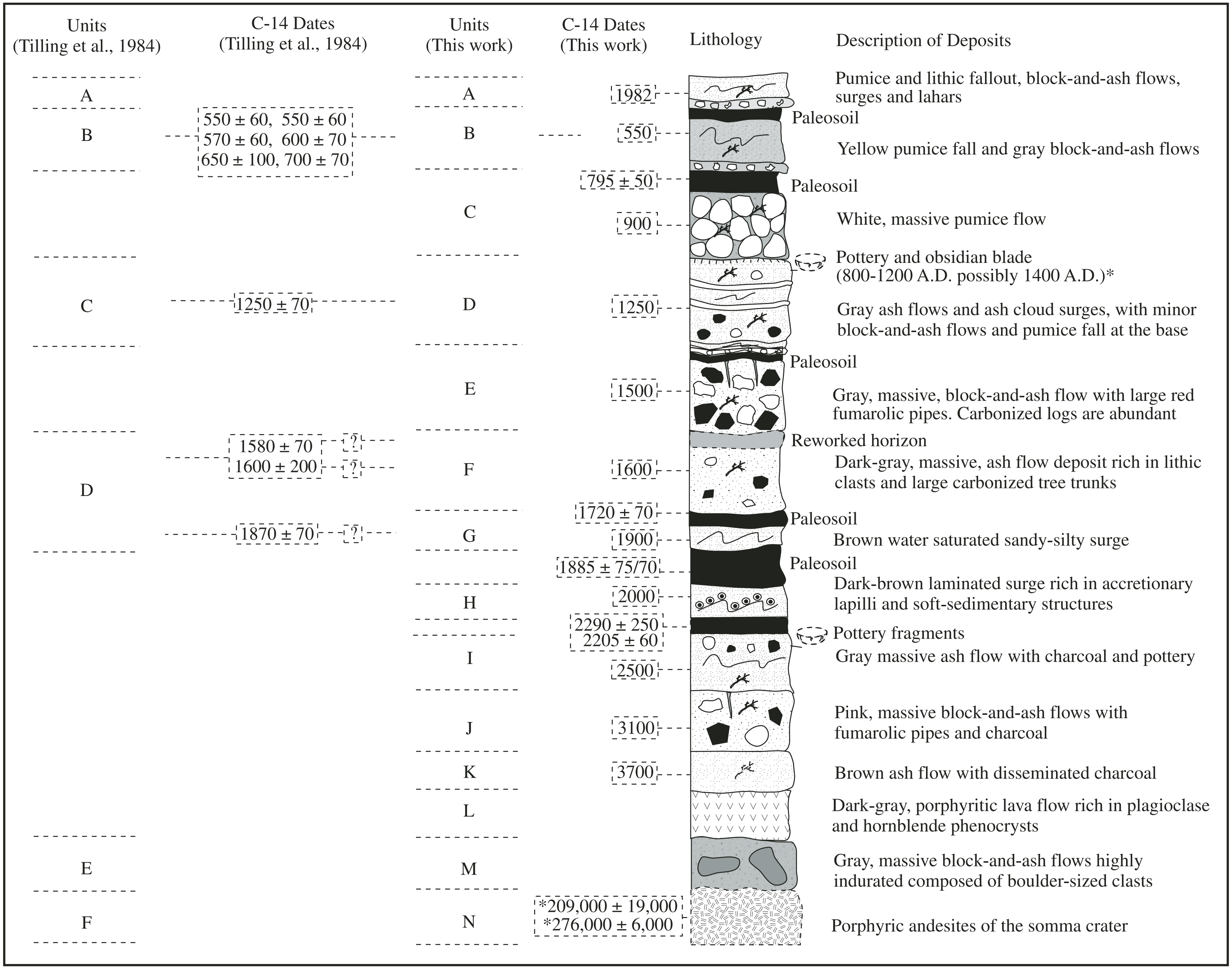


Figure 27. Simplified stratigraphic column of El Chichón volcano with the deposits so far described during the Holocene (from Espíndola et al., 2000).

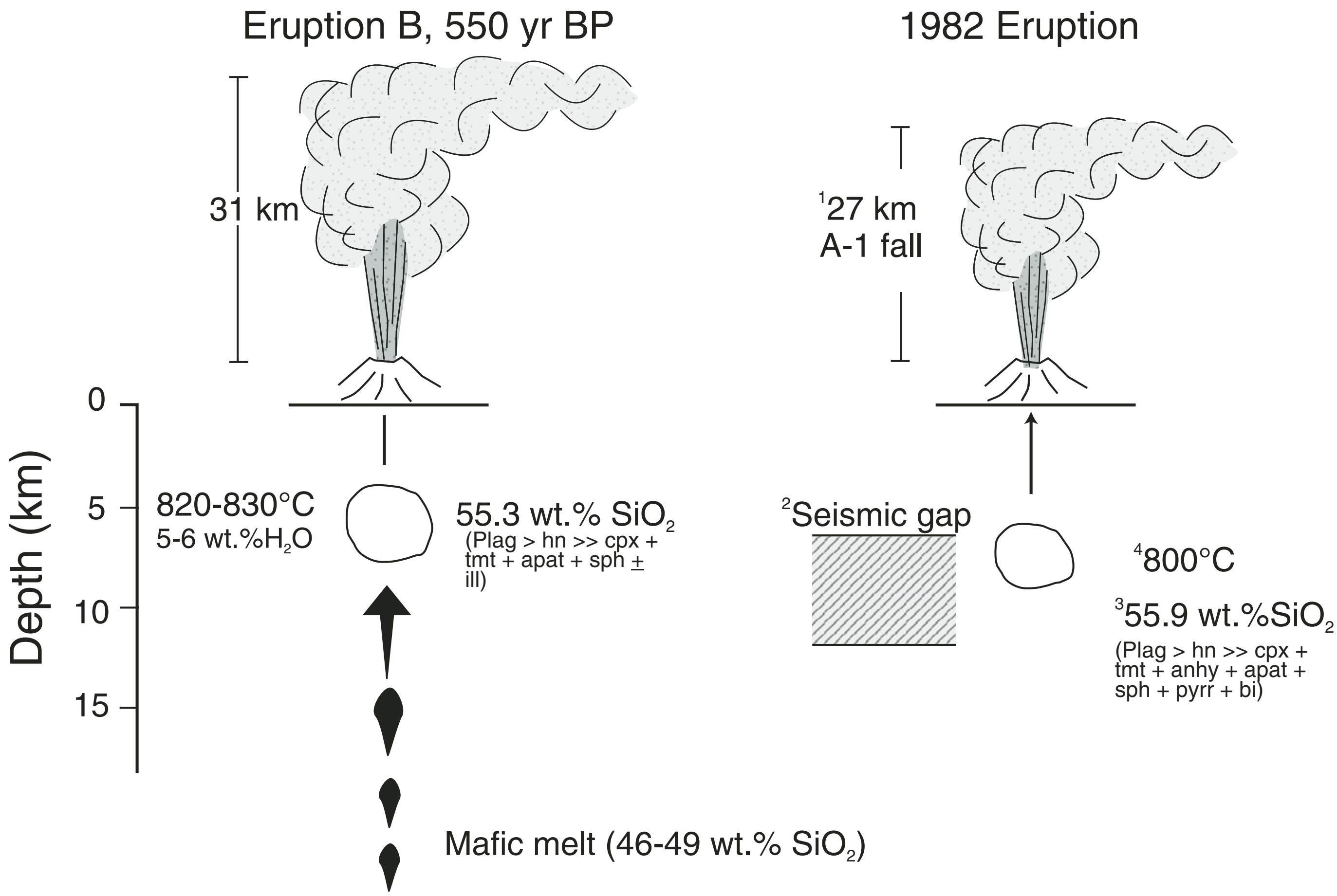


Figure 28. Sketch with the location of the magma chamber beneath El Chichón volcano during the 500 yr B.P. (Macías et al., 2003) and 1982 eruptions (Jiménez et al., 1999; Luhr, 1990).

cluded that Soconusco and Istak were synonyms of Tacaná (De Cserna et al., 1988). The first general descriptions of Tacaná were made by Sapper (1896, 1899). Bergeat (1894) made the first petrographic description of the Tacaná rocks, which he described as augite andesites. Detailed studies of the volcano began after a series of earthquakes on 22 September 1902 and the 24 October 1902 eruption of Santa Maria volcano in Guatemala. These events captured the attention of several geologists who visited the region at that time. The first detailed account of Tacaná was made by Böse (1902, 1903), who reported that the volcano base was located at an elevation of 2200 masl on top of a granitic basement. Böse also concluded that Tacaná consisted of three terraces located at elevations of 3448, 3655, and 3872 m (the uppermost crater hosted by a lava dome). The rocks collected by Böse were then petrographically analyzed by Ordóñez, who classified them as hypersthene-hornblende andesites. Böse (1902, 1903) also described an explosion crater with an elliptical shape 50 m in diameter and 5 m deep, located SW of the main summit, from which sulfur waters and fumes emanated at that time. Sapper (1897) pointed out that after the 12 January 1855 earthquake, there was formation of fissures on the flanks of Tacaná, with brief emission of fumes (Mooser et al., 1958). This activity might have been related to accounts by local inhabitants, who described an eruption at the volcano summit that ejected ash from fan-shaped holes and may account for another eruption in 1878. Böse (1902, 1903) and Waibel (1933) considered Tacaná as a dormant but not extinct, volcano.

The 1949 Eruption

On 22 September 1949 an earthquake took place at Tacaná volcano, after which the residents observed white columns and ash fall on the outskirts of Unión Juárez. This event alerted the local authorities, who asked for help from the Geology Institute of the Universidad Nacional Autónoma de México (UNAM). In January 1950 the geologist Müllerried visited the area and prepared a detailed account of the eruption (Müllerried, 1951).

Mülleried observed that the Tacaná summit crater was 70 m below the volcano summit (third step of Böse, 1902) and that it was open to the N-NW. He also described the other two steps described by Böse (1902): one located 160 m below the sum-

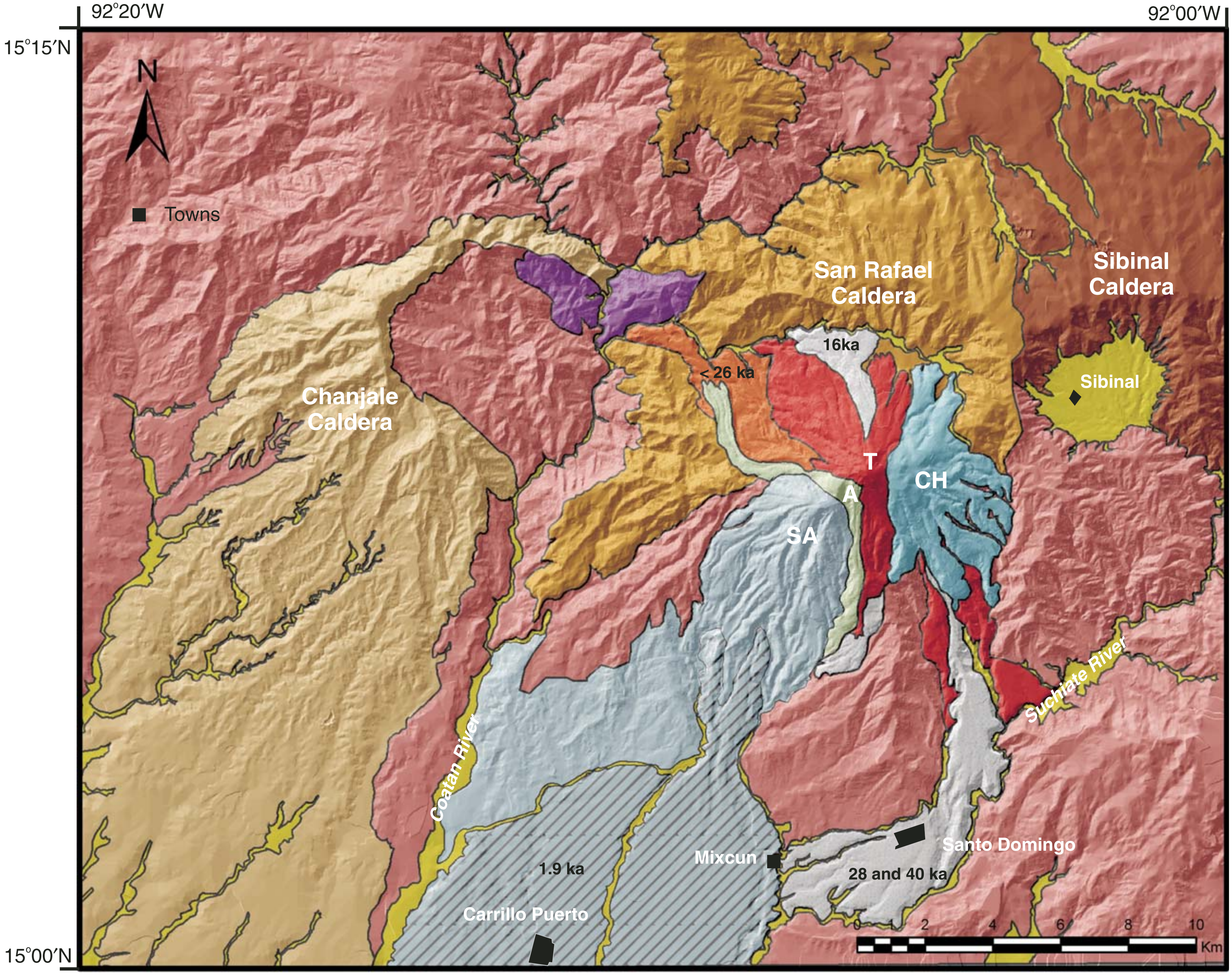


Figure 29 (*legend on following page*). General geological map of the Tacaná Volcanic Complex that is hosted inside the remains of the 1 Ma San Rafael caldera. The map shows the extent of some deposits younger than 50,000 yr B.P. from Tacaná volcano and the 1950 yr B.P. Mixcun deposits (after García-Palomo et al., 2006).

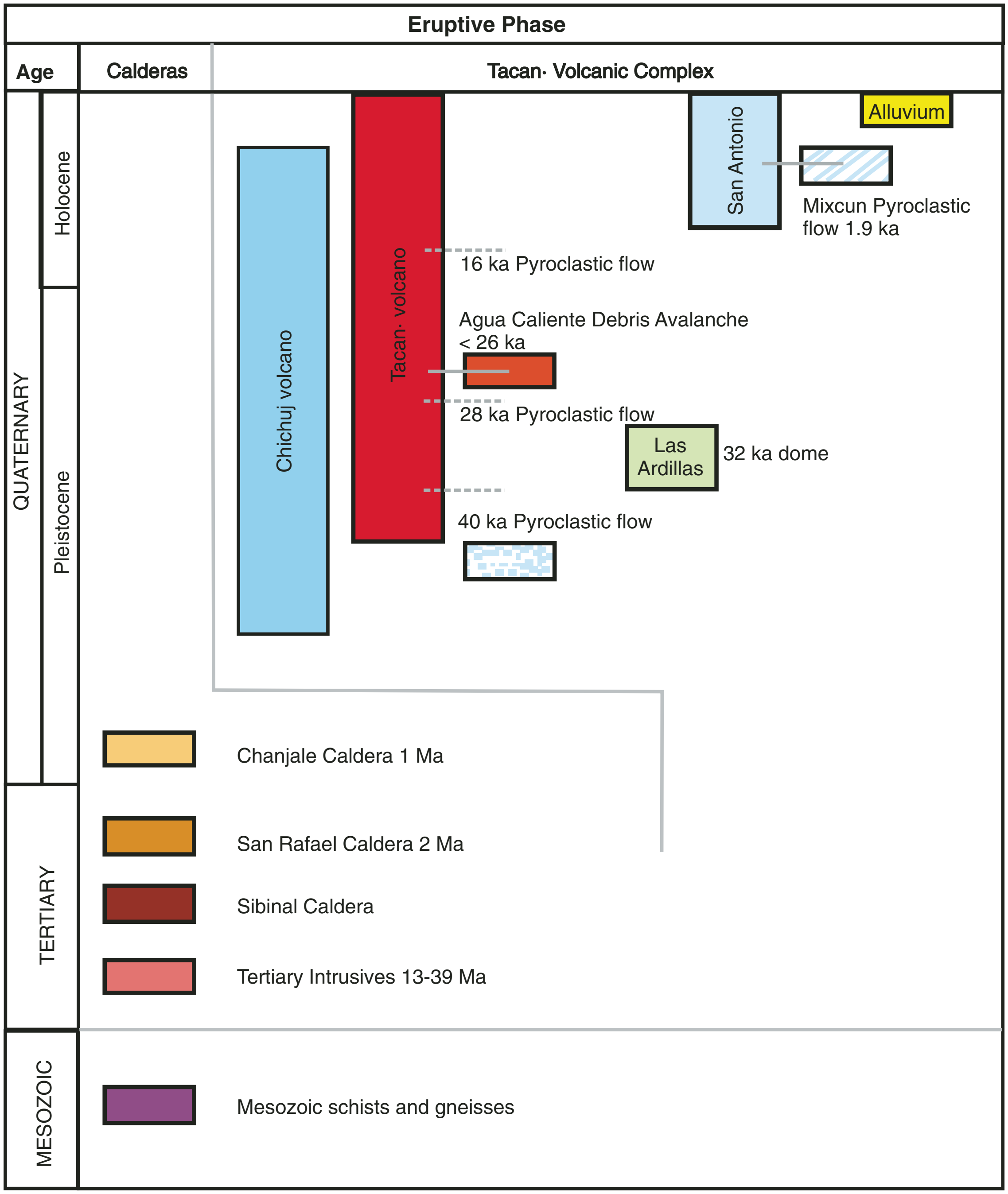


Figure 29 (*legend*).

mit with a dry lake, and the other 230 m below the summit at the lake site. According to Müllerried (1951), the 1949 activity occurred at 16 vents situated in three areas southwest of the summit. The area of these vents coincides with the suspected area of the nineteenth century eruptions mentioned by Böse (1902, 1903). The 1949 eruptive vents were 2–4 m wide and up to 4 m deep (fumarole 1; Müllerried, 1951). White to transparent fumes escaped from all the vents and were visible from the City of Tapachula. Some of these fumaroles emitted sulphurous gases and sulphur, and chlorine precipitations at their vents. The 1949 eruption pushed Tacaná into the catalogue of active volcanoes of the world (Mooser et al., 1958).

The 1986 Eruption

On December 1985, after 35 yr of repose, Tacaná volcano awoke with earthquakes and noises that continued thorough January 1986, when a seismic network was deployed at the volcano.

The most important seismic event occurred on 3 February 1986, producing damage to adobe homes at the Ixchiguan village in the San Marcos Department, Guatemala, located 25 km ENE from the crater. The seismic activity increased until 7 May, when an earthquake swarm provoked panic among the population. On 8 May, when the frequency of “felt” earthquakes by the population was two per minute, a phreatic explosion occurred outside the summit crater, almost along the international border, at an elevation of 3600 m (De la Cruz-Reyna et al., 1989). The explosion produced an 8-m-wide vent from which rose a 1-km-high column rich in water vapor and gases. Afterward, the seismic activity declined notably, and two days later this activity reached levels such as those recorded in April 1986. The fumarole was enriched in water vapor, without magmatic components (Martíni et al., 1987).

Prior to this eruption, CFE had begun a series of studies to evaluate the geothermic potential of the volcano using geological reconnaissance (Medina, 1985; De-la Cruz and Hernández, 1985) and the chemistry of the water springs around the volcano (Medina-Martínez, 1986). The latter pointed out that the spring water of Tacaná contains acid sulfate and reported fumaroles located 3200 and 3600 m to the S-SW from the summit, with temperatures of 82 and 94 °C. Today, the spring waters of Tacaná are rich in CO_2, with a composition that can be interpreted as a mixture of a deep source rich in SO_4-HCO_3-Cl and dilute meteoric water (Rouwet et al., 2004). The total emission of volatiles at Tacaná is ~50 t/d of SO_2, a typical value for passive degassing volcanoes.

De la Cruz and Hernández (1985) made the first geological map of the volcano at a scale of 1:120,000 and constructed the first stratigraphic column of the area with the volcano sitting atop a granitic basement and Tertiary andesites. In addition, these authors and Saucedo and Esquivias (1988) mapped three pyroclastic fans, dubbed Qt1, Qt2, and Qt3, which they associated with the formation of three calderas. Later, De Cserna et al. (1988) presented a photogeological map of the volcano at a 1:50,000 scale, defined 14 stratigraphic units, and summarized previous studies. In their study, these authors concluded that Tacaná is a polygenetic stratovolcano composed of three NE-SW aligned volcanoes, formed during the eruptive periods named Talquian, Tacaná, and El Águila. It is likely that these volcanoes correspond to the three calderas proposed by De la Cruz and Hernández (1985). Similarly, Mercado and Rose (1992) produced a photogeological map of Tacaná, hazard maps for the different types of hazards, and defined the chemistry of the rocks as calc-alkaline andesites. Recent studies have pointed out that Tacaná is actually a volcanic complex composed of four aligned structures in a NE-SW direction, as suggested by De Cserna et al. (1988), whose activity migrated from NW to the SE. These structures are, from oldest to youngest: Chichuj volcano (Talquian; De Cserna et al., 1988), Tacaná (Tacaná; De Cserna et al., 1988), las Ardillas dome, and San Antonio (El Águila: De Cserna et al., 1988; Macías et al., 2000; García-Palomo et al., 2006) (Fig. 30).

Regional Geology

Tacaná is located close to the triple point junction between the North America, Caribbean, and Cocos plates (Burkart and Self, 1985; Guzmán-Speziale et al., 1989). The boundary between the North America and Caribbean plates is given by the left lateral Motagua-Polochic Fault System, which separates the Maya block to the north and the Chortis block to the south (Ortega-Gutiérrez et al., 2004), where Tacaná volcano is located (Fig. 29). Previous studies suggest that Tacaná was built upon a Paleozoic basement (Mooser et al., 1958; De Cserna et al., 1988; De la Cruz and Hernández, 1985). However, due to its tectonic location within the Chortis block, its basement rocks, represented by schists and gneisses, should be younger, of Mesozoic age (García-Palomo et al., 2006). The Tacaná area was then affected by two intrusion episodes of granites, granodiorites, and tonalities, occurring at 29–35 Ma and 13–20 Ma (Mugica, 1987; García-Palomo et al., 2006). The volcanic activity began 2 Ma with the formation of

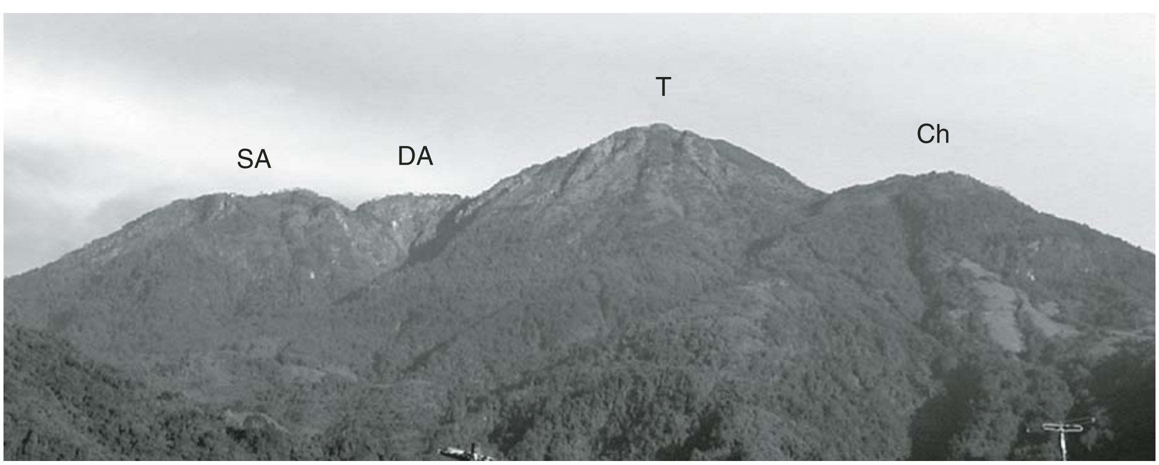


Figure 30. Panoramic view to the north of the Tacaná Volcanic Complex from the Unión Juárez village. The edifices are Chichuj (Ch), Tacaná (T), Domo las Ardillas (DA), and San Antonio (SA).

the San Rafael caldera and continued 1 Ma with the formation of the Chanjale caldera and the Sibinal caldera (García-Palomo et al., 2006). The Tacaná Volcanic Complex began its construction ca. 100,000 yr B.P. inside the San Rafael caldera.

Eruptive History

The first radiometric dates in charcoal of Tacaná volcano were obtained by Espíndola et al. (1989). These authors dated a block-and-ash flow deposit exposed at La Trinidad village at 42,000 yr B.P. This deposit corresponded to the oldest Qt3 pyroclastic fan of De la Cruz and Hernández (1985). Based on this age, Espíndola et al. (1989) considered that these deposits could correspond to some of the oldest deposits of Tacaná. Later on, Espíndola et al. (1993) obtained another radiometric age for the same deposit, yielding an age of 38,000 yr B.P., which confirmed the age of the deposit. In addition, these authors described another block-and-ash flow deposit near the Monte Perla locality that gave a date of 30,000 yr B.P.; this deposit was exposed within the pyroclastic fan Qt3. This new age suggested that the pyroclastic fan Qt3 consisted of several deposits associated with eruptions produced at Tacaná volcano, the tallest structure of the Tacaná Volcanic Complex. Later, Macías et al. (2000) identified deposits related to two eruptions dated at ca. 10,000 and 1950 yr B.P. The 1950 eruption, however, was originated at San Antonio volcano, the youngest volcano of the Tacaná Volcanic Complex. This eruption produced a later explosion that destroyed the summit dome, generating some pyroclastic surges but also a block-and-ash flow deposit that traveled up to 14 km from the summit along the Cahuacán and Mixcun ravines. The deposit is well exposed on the outskirts of the Mixcun village from which it takes its name (Macías et al., 2000) and corresponds to pyroclastic fan Qt2 (De la Cruz and Hernández, 1985). At that time, the main cultural center of the region was Izapa, a ceremonial center located ~17 km from the Tacaná summit. The archaeological excavations at Izapa have suggested that the center had been temporarily abandoned during the first century A.D. (Lowe et al., 1982). In this part of the excavation, Lowe et al. found objects with features similar to pottery produced in Guatemala and El Salvador, a fact that led these authors to propose that villagers of Izapa abandoned the city to conquer such lands. However, the 1950 yr B.P. eruption of San Antonio Volcano was dated within the period of the abandonment at Izapa (Macías et al., 2000) defined by Lowe et al. (1982), suggesting that the cause of the abandonment was the eruption of Tacaná itself. The pyroclastic flow produced by the 1950 yr B.P. eruption did not directly impacted Izapa; however, secondary hyperconcentrated and debris flows caused extensive damage in the area, destroying crops and isolating Izapa from Central México (Macías et al., 2000). This phenomenon would have caused the villagers of Izapa to migrate to Central America. This phenomenon is not uncommon in Chiapas, since tropical storms and hurricanes have isolated Tapachula from México and Central America during the last two decades.

The eruptions that occurred at 40,000, 30,000, and 1950 yr B.P. caused the partial or total destruction of andesitic central domes, and this generated block-and-ash flows able to move and infill ravines as far as 15 km around the volcano. Recent studies have identified at least other six eruptions, occurring at 32,000, 28,000, <26,000, 16,000, 7500, and 6500 yr B.P. (Macías et al., 2004b; Mora et al., 2004; García-Palomo et al., 2006), in addition to the 40,000, 30,000, 10,000, and 1950 yr B.P. eruptions. Some of these eruptions have not been considered in previous hazard zonation studies of the volcano (Mercado and Rose, 1992; Macías et al., 2000). The eruptions that occurred 40,000, 30,000, 16,000 and 1950 yr B.P. were produced by the partial destruction of andesitic domes from Tacaná and San Antonio volcanoes.

The 32,000 yr B.P. eruption formed a Plinian column that was dispersed to the NE into the Guatemala lands and that deposited a pumice fall layer. Prior to this report, there was a brief note of the presence of 2-m-thick fall deposits close to the Sibinal village, Guatemala, by Mercado and Rose (1992).

The 26,000 yr B.P. eruption was caused by the collapse of the NW part of Tacaná volcano. At that time, the crater was occupied by an andesitic central dome. The collapse produced a debris avalanche followed by a series of block-and-ash flows that traveled at least 8 km to the Coatán River. The debris avalanche has an H/L of 0.35, covers a minimum area of 8 km^2, and has a volume of 1 km^3 (Macías et al., 2004b; García-Palomo et al., 2006). The eruption ended with the generation of debris flows along the San Rafael and Chocab rivers.

The eruptions dated at ca. 7500 and 6500 yr B.P. produced ash flows and pyroclastic surges, respectively, forming a thin blanket that covers the Tacaná cone.

Tacaná volcano has mainly erupted two pyroxene andesites and minor dacites. The andesites are composed of plagioclase, augite, hypersthene + FeO oxides, and rare hornblende. The chemical composition of these rocks is very homogeneous (Mercado and Rose, 1992; Macías et al., 2000; Mora, 2001; Mora et al., 2004). Therefore, Chichuj volcano has been edified by andesitic lava flows and domes (59–63 wt% SiO_2), Tacaná volcano by rare basaltic andesite enclaves (56–61 wt% SiO_2), andesitic lava flows and dacitic domes (61–64 wt% SiO_2), and pyroclastic flows with juvenile andesitic lithics (60–63 wt% SiO_2). San Antonio volcano has formed andesitic lava flows and dacitic domes (58–64 wt% SiO_2).

DISCUSSION AND CONCLUSIONS

Most of the large volcanoes in México are located along the Trans-Mexican Volcanic Belt, a continental arc that transects Central México between the 19° and 20° parallel latitude north (Fig. 1). Volcanism also appears concentrated in small areas, such as the Tres Vírgenes Volcanic Complex and the San Quintín Volcanic Field in Baja California, the Pinacate Volcanic Field in Sonora, the Revillagigedo archipelago in the Pacific, Los Tuxtlas Volcanic Field in Veracruz, the Chiapanecan Volcanic Arc, and the northwestern edge of the Central America Volcanic Arc in

Chiapas. Several authors have concluded that the Trans-Mexican Volcanic Belt is formed by the subduction of the Rivera and Cocos beneath the North America plate at the Middle American Trench (Ponce et al., 1992; Singh and Pardo, 1993; Pardo and Suárez, 1993, 1995). The Trans-Mexican Volcanic Belt has an oblique position of ~15° with respect to the Middle American Trench; this is a strange feature for continental volcanic arcs. This feature caused some authors to propose that the Trans-Mexican Volcanic Belt is related to a megashear transecting the central part of México (Cebull and Shurbet, 1987) or to active extension like a rift zone (Sheth et al., 2000).

The Central America Volcanic Arc extends from western Panama to southeastern México. It runs parallel to the Middle America Trench, and at the México-Guatemala border volcanism becomes discontinuous and migrates inland along with the seismic isodepth curves (Fig. 31). The discontinuous volcanism appears as a scatter of landforms in the state of Chiapas forming the Chiapanecan Volcanic Arc, which has mainly erupted calc-alkaline products but also K-rich alkaline rocks at el Chichón, located 350 km from the trench, and Na-rich alkaline products at Los Tuxtlas Volcanic Field, located 400 km from the trench. Nixon (1982) proposed that the alkaline volcanism of El Chichón and Los Tuxtlas was due to extensional tectonics related to the triple point junction of the North America, Caribbean, and Cocos plates. However, other authors have concluded that the volcanism of the region is related to the subduction of the Cocos plate beneath the North America plate (Stoiber and Carr, 1973; Thorpe, 1977; Burbach et al., 1984; Bevis and Isacks, 1984; Luhr et al., 1984; García-Palomo et al., 2004).

Finally, the volcanism reappears in Central México at parallel 19°N at the Trans-Mexican Volcanic Belt, extending from the coast of Veracruz to the coast of Nayarit predominantly as calc-alkaline volcanism but with small regions of alkaline volcanism, such as the Colima graben and the Chichinautzin Volcanic Field (Fig. 31). The thickness of the continental crust increases from west to east on the Trans-Mexican Volcanic Belt: at Colima, the crust is 20–22 km thick; at Nevado de Toluca, it is 40 km thick; the crust at Popocatépetl is 47 km thick; and at Pico de Orizaba, it is >50 km thick (Molina-Garza and Urrutia-Fucugauchi, 1993; Urrutia-Fucugauchi and Flores-Ruiz, 1996). These variations are reflected in the $^{87}Sr/^{86}Sr$ versus εNd isotopic relationships of the volcanoes (Nelson et al., 1995; Macías et al., 2003; Martínez-Serrano et al., 2004; Schaaf et al., 2004) (Fig. 32). Colima is the closest volcano to the trench with the most primitive isotopic relationships compared to Nevado de Toluca, Popocatépetl, and Pico de Orizaba, which are located farthest from the trench with a thicker continental crust below them. Because of to this, the volcanoes show clear evidence of crustal contamination, as observed through Osmium isotopes (Chesley et al., 2000; Lassiter and Luhr, 2001) and the presence of crustal xenoliths (Valdez-

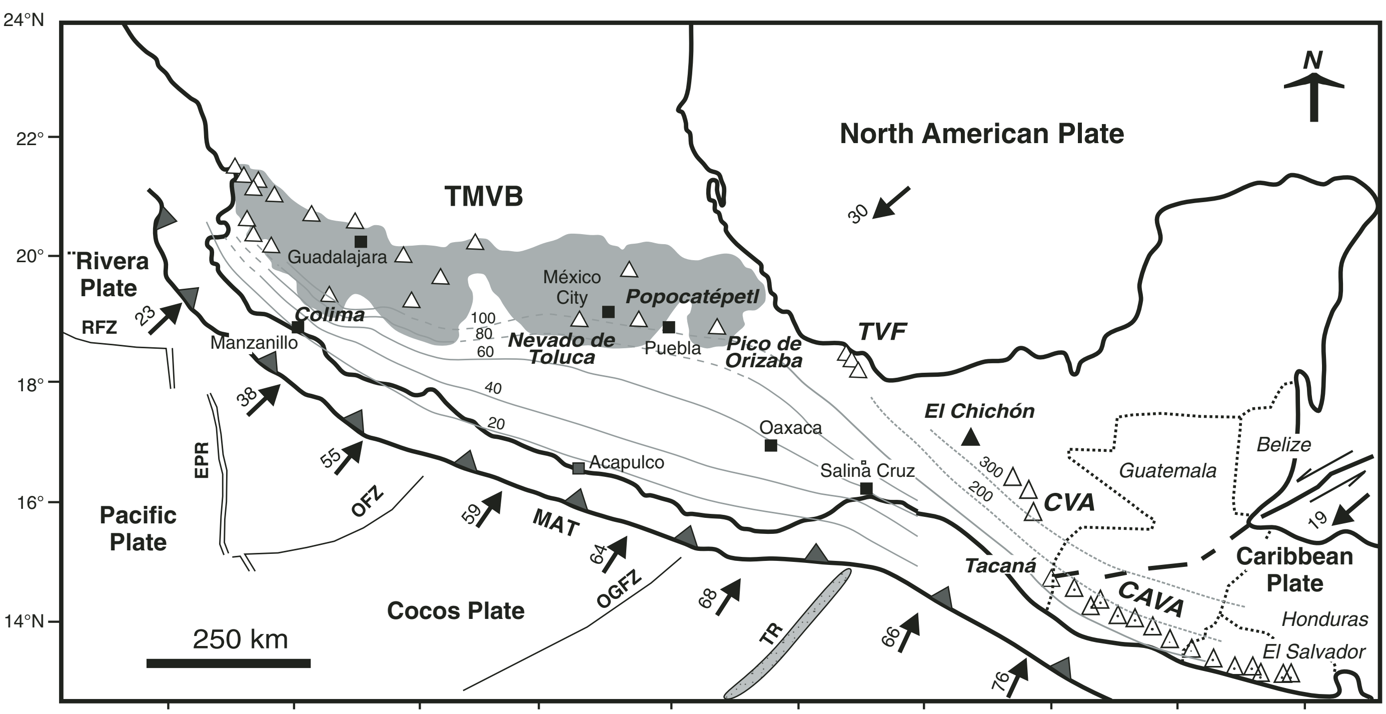


Figure 31. Tectonic setting of the southern portion of México with the location of the Trans-Mexican Volcanic Belt (TMVB), Los Tuxtlas Volcanic Field (TVF), Chiapanecan Volcanic Arc (CVA), and the Central America Volcanic Arc (CAVA) with respect to the plates. The gray lines are the isodepth curves of earthquakes. EPR—East Pacific Rise; MAT—Middle America Trench; OFZ—Orozco Fracture Zone; OGFZ—O'Gorman Fracture Zone; RFZ—Rivera Fracture Zone; TR—Tehuantepec Ridge (after García-Palomo et al., 2004).

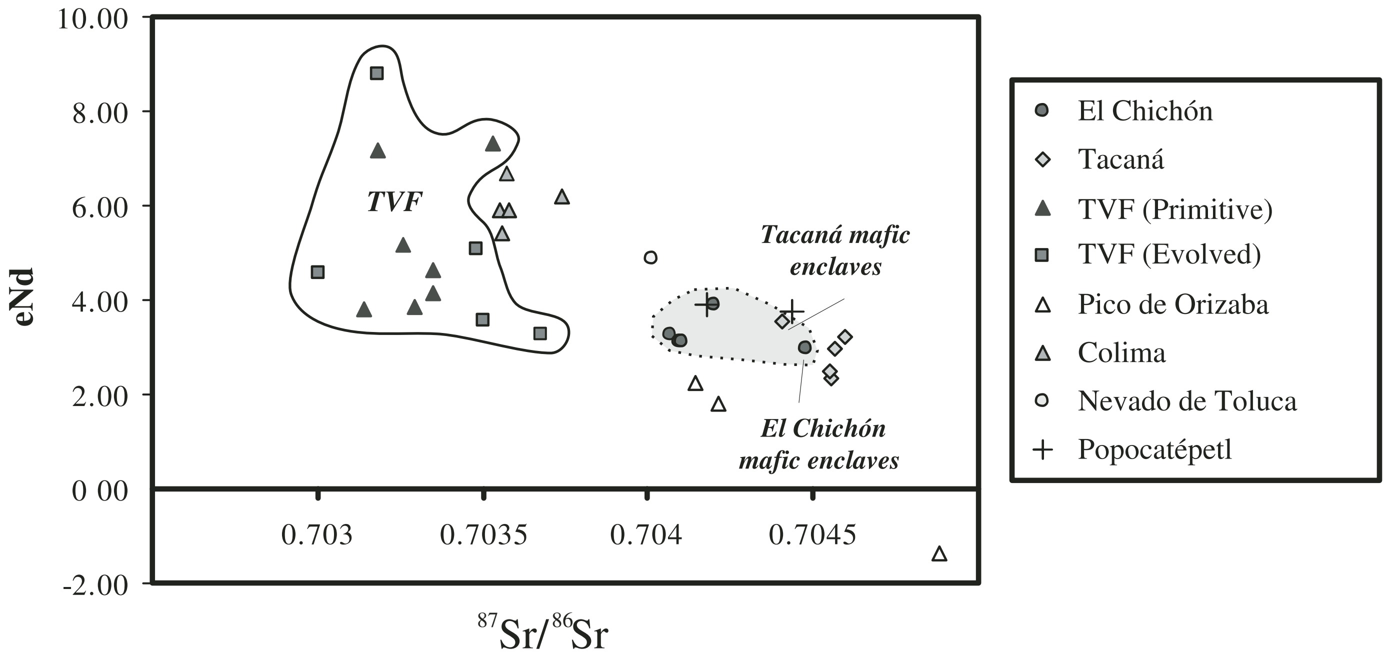


Figure 32. Isotopic ratios of $^{87}Sr/^{86}Sr$ versus epsilon neodymium of rocks from Colima, Nevado de Toluca, Popocatépetl, Pico de Orizaba, Tuxtlas Volcanic Field (TVF), Chichón, and Tacaná volcanoes.

Moreno et al., 2006). The degree of crustal assimilation is related to the age and composition of the continental crust through which these magmas interact as well as the storage time at the base of the crust and their subsequent ascent to the surface. Schaaf et al. (2004) concluded that the primitive isotopic relationships of Colima volcano are due to the presence of a young and primitive continental crust while the more evolved isotopic relationships of Popocatépetl and Pico de Orizaba are due to the presence of the Morelos Formation and Grenvillian rocks, respectively.

In southeastern México, there are volcanic products with different characteristics (Fig. 32). Tacaná volcano is located 100 km from the trench, atop a 40-km-thick continental crust (Rebollar et al., 1999). This volcano has isotopic ratios with a higher degree of crustal assimilation ($^{87}Sr/^{86}Sr$: 0.70441–0.70459; εNd: 2.26–3.57) (Mora et al., 2004), with respect to the isotopic ratios of Chichón volcano (Macías et al., 2003), located 250 km from the trench, and several volcanoes of Los Tuxtlas Volcanic Field, located 350 km from the trench (Nelson et al., 1995). The volcanic rocks erupted in southern México have a calc-alkaline signature with medium K contents, negative anomalies of Nb, Ti, and P, and enrichments in light rare earth elements, typical of subduction environments.

The compositional features (major, trace, rare earth elements, some isotope ratios) of Mexican volcanoes have been established during the past 30 yr. Despite these advances, we still known very little about the processes of magma genesis, the partial fusion of the mantle that occurred at depths of 75 km or deeper, the assimilation and contamination processes at depths of <50 km, magma storage, and other processes (fractional crystallization, magma mixing, etc.) occurring at shallower depths. There have been some recent advances in experimental petrology studies of our volcanoes attempting to understand the pre-eruptive conditions of magmas (i.e., pressure, temperature, and water saturation) of some eruptions. For instance, the magma erupted during the 1998 eruption of Colima volcano had a pre-eruptive temperature of 840–900 °C at depths of 3–7 km below the crater (Mora et al., 2002). During the April 1996 through February 1998 eruptions at Popocatépetl (Straub and Martín-Del Pozzo, 2001), the volcano erupted a hybrid magma originated by the mixing of an olivine-spinel–saturated andesitic magma (55% SiO_2, 1170–1085 °C) probably from the Moho and a clynopyroxene-orthopyroxene-plagioclase dacitic magma (62% SiO_2, ~950 °C) stored between ~4 and 13 km below the crater. Between 14,000 and 10,500 yr B.P., three eruptions occurred at Nevado de Toluca volcano that erupted magmas with the same chemical and mineral association that Arce et al. (2005a) interpreted as the same magma stagnated at a temperature of 814–840 °C and at depths of 6 km beneath the crater. Wallace and Carmichael (1999) concluded that some olivine basalts at the Chichinautzin Volcanic Field had temperatures of 1200–1290 °C prior to eruption. In the eastern Trans-Mexican Volcanic Belt, recent lavas erupted at two vents of Volcancito had pre-eruptive temperatures of 1198 and 1236 °C (calc-alkaline basalt Rio Nolinco) and 1166–1175 °C (hawaiite Toxtlacuaya) (Carrasco-Núñez et al., 2005). All these studies represent a considerable advance in our knowledge of shallow processes; however, our knowledge of deeper processes is still lacking and may improve with the study of mantle xenoliths (Blatter and Carmichael, 1998) and fluid inclusions (gas and melt inclusions) trapped in the magmas. There are already some studies in this field, such as those accomplished for Parícutin volcano (Luhr, 2001) and the Chichinautzin Volcanic Field (Cervantes and Wallace, 2003).

Other studies that could allow better understanding of the magmatic processes use Boron isotopes to estimate the influence of sediments in the subduction process (Hochstaedter et al., 1996), hydrogen isotopes to understand the isotopic characteristics of the magmatic water (Taran et al., 2002), and Sr-Nd-Pb isotopic ratios to evaluate petrologic process and the source of the magmas (Martínez-Serrano et al., 2004; Schaaf et al., 2005; Valdéz-Moreno et al., 2006).

As stated in previous sections, there has been a considerable advance in the study of the stratigraphic record of the Mexican volcanoes, which is necessary to better comprehend their eruptive history and to design hazard maps. In most Mexican volcanoes, the stratigraphic record dates back to 50,000 yr B.P. as covered by the radiocarbon dating method. With this period of time, we can document the eruptive record of Paleofuego and Colima volcano but not the stratigraphic record of Nevado de Colima. In other words, most of our volcanoes began their formation beyond the ^{14}C limit ca. 1–2 Ma. In order to completely understand the eruptive behavior of these volcanoes, other radiometric methods are needed, such as K-Ar, Ar-Ar in minerals and rocks, thermoluminiscense in minerals, ^{36}Cl in polished surfaces of moraine deposits, etc.

Geologic studies (petrology, geochemistry, volcanology, etc.) and the historic record are very important to understanding the past behavior of our active volcanoes and to forecast future activity. For instance, the eruptive record of Popocatépetl shows that during the past 23,000 yr, the volcano has had four major Plinian eruptions and that, therefore, a large future event might be of this type. The eruptive records of Paleofuego and Colima volcanoes indicate that these cones have produced debris avalanches and many small dome collapse events. Minor activity at Colima will generated small-volume pyroclastic flows like those produced during the past 20 yr. It is clear that based upon the eruptive record of a volcano, we must construct hazard maps that show the areas affected by eruptions with different magnitudes in the past. To date, México has hazard maps of Colima, Popocatépetl, and Pico de Orizaba volcanoes and preliminary hazard maps of Tres Vírgenes, Nevado de Toluca, Chichón, and Tacaná. These maps represent a basic source of information for the Civil Protection authorities, who should use them to create preventive emergency plans, education programs to the population, etc. (Macías and Capra, 2005).

The integration of the geologic information (eruptive history, historic record, hazard maps, chemistry of the erupted products, etc.) and the geophysical information (seismology, gravimetry, magnetometry, etc.) is key to understanding the present behavior of our volcanoes. The 1994–present eruption of Popocatépetl has allowed the gathering of this information in order to establish a monitoring network to forecast future events. Despite of all this development, we are not yet able to analyze and prepare a conceptual model of Popocatépetl with the depth, shape, and volume of the magma chamber, ascent times of the magma to the surface, and its relationship to degassing activity at the surface.

During the next decades, volcanologic studies in México should focus on completing the eruptive record of the volcanoes, to determine the magmatic processes that take place at depth, the storage conditions in the upper crust, the chemical evolution of the products, the times of magma ascent to the surface, and to update and create hazards maps of all the active volcanoes. Finally, all of this information should be available to the population living around the volcanoes in order to avoid a future volcanic disaster such as the 1982 eruption of El Chichón. An eruption cannot be stopped, but its consequences can be reduced considerably.

ACKNOWLEDGMENTS

This work has benefited from funding of diverse projects, among them Consejo Nacional de Ciencia y Tecnología (CONACYT, 47226) and Dirección General de Asuntos del Personal Académico de la Universidad Nacional Autónoma de México (DGAPA-UNAM, IN114206-3). Most of the observations and fieldwork carried out in these volcanoes were made with friends and colleagues: J.M. Espíndola at El Chichón and Tacaná volcanoes; C. Siebe at Popocatépetl; R. Saucedo, M.I. Bursik, J.C. Gavilanes, and A. Cortés at Colima volcano; C. Siebe and G. Carrasco at Pico de Orizaba volcano; and A. García, L. Capra, J.L. Arce, R. Saucedo, J.C. Mora, T. Scolamacchia, K. Scott, S. Hughes, and D. Sarocchi at all these volcanoes. All satellite imagery was courtesy of Michael Abrams from the Jet Propulsion Laboratory–National Aeronautics and Space Administration, USA. G. Carrasco and M. Ort provided careful review of this manuscript. I am thankful to C. Limon for producing the geologic maps of the volcanoes, C. Ordaz for looking at dark, old manuscripts in the UNAM library and putting together the reference list, Sara Solís for correcting the final version of the reference list, and A.M. Rocha for her technical support. I am grateful to several people who provided logistic support at these volcanoes, such as E. Segre and A. Fernández at Popocatépetl volcano and S. Hernández at Chichón and Tacaná volcanoes. Finally, I would like to give my most sincere appreciation to Reynaldo Mota, Servando de la Cruz, Robert I. Tilling, Jorge Neyra, R. Díaz, and G. Carrasco for the material they have provided to me during the course of the last decade.

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Manuscript Accepted by the Society 29 August 2006