Late Oligocene to Middle Miocene rifting and synextensional magmatism in the southwestern Sierra Madre Occidental, Mexico: The beginning of the Gulf of California rift

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ABSTRACT

Although Basin and Range–style extension affected large areas of western Mexico after the Late Eocene, most consider that extension in the Gulf of California region began as subduction waned and ended ca. 14–12.5 Ma. A general consensus also exists in considering Early and Middle Miocene volcanism of the Sierra Madre Occidental and Comondú Group as subduction related, whereas volcanism after ca. 12.5 Ma is extension related. Here we present a new regional geologic study of the eastern Gulf of California margin in the states of Nayarit and Sinaloa, Mexico, backed by 43 new Ar-Ar and U-Pb mineral ages, and geochemical data that document an earlier widespread phase of extension. This extension across the southern and central Gulf Extensional Province began between Late Oligocene and Early Miocene time, but was focused in the region of the future Gulf of California in the Middle Miocene. Late Oligocene to Early Miocene rocks across northern Nayarit and southern Sinaloa were affected by major approximately north-south–to-north-northwest–striking normal faults prior to ca. 21 Ma. Between ca. 21 and 11 Ma, a system of north-northwest–south-southeast high-angle extensional faults continued extending the southwestern side of the Sierra Madre Occidental. Rhyolitic domes, shallow intrusive bodies, and lesser basalts were emplaced across this extensional belt at 20–17 Ma. Rhyolitic rocks, in particular the domes and lavas, often show strong anatectic inheritance but only a few Mesozoic or older xenocrysts, suggesting silicic magma generation in the mid-upper crust triggered by an extension-induced basaltic influx. In northern Sinaloa, large grabens were occupied by huge volcanic dome complexes ca. 21–17 Ma and filled by continental sediments with interlayered basalts dated as 15–14 Ma, a stratigraphy and timing very similar to those found in central Sonora (northeastern Gulf of California margin). Early to Middle Miocene volcanism occurred thus in rift basins, and was likely associated with decompression melting of upper mantle (inducing crustal partial melting) rather than with fluxing by fluids from the young and slow subducting microplates. Along the eastern side of the Gulf of California, from Farallón de San Ignacio Island offshore Los Mochis, Sinaloa, to San Blas, Nayarit, a strike distance of >700 km, flat-lying basaltic lavas dated as ca. 11.5–10 Ma are exposed just above the present sea level. Here crustal thickness is almost half that in the unextended core of the adjacent Sierra Madre Occidental, implying significant lithosphere stretching before ca. 11 Ma. This mafic pulse, with subducted Nb-Ta negative spikes, may be related to the detachment of the lower part of the subducted slab, allowing an upward asthenospheric flow into an upper mantle previously modified by fluid fluxes related to past subduction. Widespread eruption of very uniform oceanic island basalt–like lavas occurred by the late Pleistocene and Pleistocene, only 20 m.y. after the onset of rifting and ~9 m.y. after the end of subduction, implying that pre-existing subduction-modified mantle had now become isolated from melt source regions. Our study shows that rifting across the southern-central Gulf Extensional Province began much earlier than the Late Miocene and provided a fundamental control on the style and composition of volcanism from at least 30 Ma. We envision a sustained period of lithospheric stretching and magmatism during which the pace and breadth of extension changed ca. 20–18 Ma to be narrower, and again after ca. 12.5 Ma, when the kinematics of rifting became more oblique.

INTRODUCTION

Over the past 30 m.y., the western North American plate margin has changed from convergence to highly oblique divergence through a complex interaction with the Pacific plate that produced two pairs of parallel structures: the San Andreas fault system and the Eastern California Shear Zone–Walker Lane in the U.S., and the San Benito–Tosco–Abreojos fault system and the Gulf of California rift in Mexico (Fig. 1). Strike-slip deformation in both regions moved inland; currently, the Eastern California Shear Zone–Walker Lane accommodates ~25% and the Gulf of California accommodates ~90% of the Pacific–North America relative plate motion (Wesnousky, 2005; Plattner et al., 2009). Stretching of the continental lithosphere preceded direct interaction of the Pacific and North America plates and produced the Basin and Range composite extensional province (Dickinson, 2002). In Mexico, the Basin and Range...
Figure 1. Regional tectonic map of the Gulf of California and adjoining areas showing the different types of lithosphere, extents of the Basin and Range and Gulf Extensional Provinces (dashed orange lines), and preserved extent of the contiguous part of the Sierra Madre Occidental silicic large igneous province (dashed red line). The extent of the unextended core of the Sierra Madre Occidental has been revised according to this study. Patterns of oceanic plate boundaries and crustal isochrons are from Lonsdale (1991) and Tian et al. (2011). GEP—Gulf Extensional Province, where different eastern boundaries in Sonora are based on those defined by Stock and Hodges (1989) and Calmus et al. (2010); MGE—Main Gulf Escarpment; BG—Bahía Guadalupe; BA—Bahía de Los Angeles; IT—Isla Tiburón; TB—Tiburón Basin; ITM—Islas Tres Marias; MF—Magdalena Fan; EPR—East Pacific Rise; SM—Sierra El Mayor. Inset shows the main fault systems currently defining the Pacific–North America plate boundary (SAF—San Andreas fault system; WL-ECSZ—Walker Lane–Eastern California shear zone; SBTA—San Benito–Tosco Abreojos fault system; GoC—Gulf of California fault system) and the Sierra Nevada (SN) and Baja California (BC) microplates; RP—Rivera Plate; JF—Juan de Fuca Plate.
Early extension in the Gulf of California

episode of extension is recognized across a wide region north of the Trans-Mexican Volcanic Belt (e.g., Henry, 1989; Henry and Aranda-Gómez, 1992, 2000; Calmus et al., 2010) (Fig. 1), although the western limit of this extension has been poorly defined. More focused extension in the Gulf of California area, bounded to the west by the Main Gulf Escarpment of Baja California and to the east by the unextended core of the Sierran Madre Occidental (SMO), has been referred to as the Gulf Extension Province (GEP) (Gastil et al., 1975) (Fig. 1) or the proto-Gulf of California (Kari and Jentsky, 1972). In earlier studies (e.g., Stewart, 1978; Henry, 1989), it was argued that proto-Gulf of California rifting substantially preceded the formation of a Pacific-North America plate boundary at this latitude (ca. 14–12.5 Ma); i.e., extension occurred when subduction was still ongoing. However, after the influential work of Stock and Hodges (1989), most (e.g., Henry and Aranda-Gómez, 1992, 2000; Umhoefer et al., 2001; Umhoefer, 2011; Fletcher et al., 2007; Lizarralde et al., 2007; Seiler et al., 2011; Sutherland et al., 2012; Miller and Lizarralde, 2013) assumed that extension in the Gulf of California region began only at the end of Middle Miocene time, when subduction ended and the transfer of Baja California to the Pacific plate began. In the past two decades workers have debated the onset of oblique rifting in the Gulf of California (e.g., Stock and Hodges, 1989; Gans, 1997; Oskin and Stock, 2003; Fletcher et al., 2007; Miller and Lizarralde, 2013), but a general consensus has existed where gulf opening was fundamentally a post-subduction event, controlled by the highly oblique, northwestward motion of Baja California that was able to rift the continental lithosphere in ~6–10 m.y. (Umhoefer, 2011).

Separating young extension of the GEP from that of the older Basin and Range is particularly challenging in the northern part of the SMO (Fig. 1) (see discussion in Calmus et al., 2010). In Sonora and Chihuahua, extension has affected the entire volcanically active region, and appears spatially contiguous and temporally continuous, leading to different interpretations on the location of the boundary between the two extensional provinces (Fig. 1). In western Chihuahua, extension is poorly dated but affects ignimbrites as young as ca. 29 Ma (McDowell and Mauger, 1994) and is considered to have started ca. 30 Ma on the basis of basin geochemistry (Cameron et al., 1989; McDowell and Mauger, 1994). Near the Chihuahua-Sonora border, extension has been recently established as ca. 27–25 Ma (Murray et al., 2013) by detailed geologic mapping and geochronology. In central Sonora, clastic sedimentation in extensional basins as well as development of metamorphic core complexes began almost concurrently along a ~200-km-wide zone at the end of the Oligocene and continued throughout the Middle Miocene (McDowell et al., 1997; Gans, 1997; Gonzalez-León et al., 2000; Vega-Granillo and Calmus, 2003; Nourse et al., 1994; Wong and Gans, 2003; Wong et al., 2010). Low-angle detachment faulting in the core complexes waned by 15 Ma (Wong et al., 2010), but high-angle normal faulting continued until ca. 8.5 Ma in coastal Sonora (McDowell et al., 1997; Mora-Alvarez and McDowell, 2000; Roldán-Quintana et al., 2004). Gastil et al. (1975) placed the eastern limit of the GEP in the coastal region of Sonora. Stock and Hodges (1989) extended it to the western SMO (Fig. 1) to include the belt of core complexes, the age of which was not known at that time; however, they considered that it could be also related to the Late Miocene opening of the Gulf of California. Based on the new knowledge about the ages of the core complexes, Calmus et al. (2010) placed the boundary further to the west at the longitude of Hermosillo (Fig. 1), because this was the boundary of post-12 Ma extension. In this way they reiterated the notion that rifting in the Gulf of California was a post-subduction process.

Pre-Late Miocene extension was also reported from a few locations in the northern Gulf of California, although these examples have generally been neglected because of overprinting by Late Miocene–Pleistocene deformation. In Baja California, Late Oligocene to Middle Miocene extension is reported at Bahía de Guadalupe (ca. 24–14 Ma; Axen, 2003), whereas extension at least partially concurrent with the Basin and Range of central-western Sonora is documented near the Main Gulf Escarpment of northern Baja California at Sierra el Mayor (ca. 15–10 Ma; Axen et al., 2000) and southern Sierra de Juárez (16–11 Ma; Lee et al., 1996) (Fig. 1). Micropaleontological studies of deep wells drilled in the Wagner, Consag, and Tíburón Basins (Fig. 1) also suggest the presence of shallow-marine sedimentation in some areas of the northern Gulf of California in the Middle Miocene (Helena et al., 2009). These data contrast with the geology of Isla Tíburón (Oskin and Stock, 2003; Bennett et al., 2012) (Fig. 1) that clearly indicates that marine incursion at this site is latest Miocene. Despite contrasting interpretations about the timing of marine sedimentation (see the Discussion), an early onset of extension in the northern Gulf of California region cannot be ruled out and the possibility exists that the Basin and Range and the GEP were at least partly overlapping in space.

Cenozoic magmatism preceding and accompanying the development of the Gulf of California has also been divided into two different episodes, supposedly controlled by the tectonic setting at the plate boundary. Oligocene and Early Miocene silicic to bimodal volcanism of the SMO silicic large igneous province in mainland Mexico, as well as the Middle Miocene intermediate Comandú Group in Baja California, has for many years simply been interpreted as the expression of suprasubduction arc magmatism (e.g., Sawlan and Smith, 1984; Sawlan, 1991; Hausback, 1984; Umhoefer et al., 2001), whereas the appearance of more heterogeneous magma types (e.g., adakites, Nb-enriched basalts, magnesian andesites) after 12 Ma has been associated with the development of the oblique-divergent plate boundary (see reviews in Pállares et al., 2008; Calmus et al., 2010). Implicit in this view is the idea that a given tectonic setting should be promptly and clearly reflected in magma composition. However, the use of geochemistry to track the transition from subduction to rifting has been proved inconclusive in Sonora (Till et al., 2009) and is challenged in Baja California, where calc-alkaline volcanism occurred well after subduction ended, until the Pleistocene (Martín-Barajas et al., 1995; Bigioggero et al., 1995; Schmitt et al., 2006; Calmus et al., 2010).

However, it has been shown that the dominantly silicic SMO volcanics have a strong crustal contribution (Ruiz et al., 1988, 1990; Albrecht and Goldstein, 2000; Bryan et al., 2008; this work), such that the calc-alkaline and other subduction-related signatures like Nb-Ta depletions essentially reflect the composition of the crust involved in partial melting and do not directly provide any constraints on the tectonic setting of magmatism. Rapid large-scale crustal melting during the Oligocene and Early Miocene (Ferrari et al., 2007) is atypical of modern subduction zones and indicates that the SMO cannot be considered a normal volcanic arc (Bryan et al., 2008, 2013). In a similar line of reasoning, the origin of the compositionally distinct, but volumetrically minor, intermediate volcanism of the Middle Miocene Comandú Group in southern Baja California has recently been questioned, i.e., this volcanism is more consistent with mixing and hybridization in upper crustal reservoirs promoted by ongoing extensional tectonics (Bryan et al., 2013).

Due to poor access and security reasons, the southeastern side of the Gulf of California has been comparatively less studied than the western side in Baja California. This region, comprising the state of Sinaloa, the northern part of Nayarit, and the western part of Durango, Zacatecas, and Jalisco (Fig. 2), includes both the unextended core of the SMO and the GEP (Fig. 1). A better definition of the onset of rifting
Figure 2. Tectonic map of the south-central Gulf of California and adjoining regions showing the main Neogene faults and crustal thicknesses (from Persaud et al., 2007; Lizarralde et al., 2007). Offshore samples not included in Figures 6 and 8–10 are shown here (R3—ROCA 3J 5; R24—ROCA 24J 33; D46—DANA 46a). RML—Río Mezquital lineament; TL—Tayoltita lineament; Bo—Bolaños graben; RCC—Río Chico–Canatlán graben; NC—Nayar caldera field; TC—Temoaya caldera; MC—Mesa Cacaxtla shield volcano; PVF—Pericos volcanic field; CVF—Choix volcanic field; JB—Jalisco block; TAF—Tosco-Abreojos fault system; SMF—Santa Margarita fault; FSI—Farallón de San Ignacio Island; SC—Santa Catalina Island; SR—Santa Rosalía; EGE—Eastern Guaymas evaporites; LCB—Los Cabos Block; MF—Magdalena Fan; Jal.—Jalisco State.

We present results of a regional geologic study supported by 43 new 40Ar–39Ar and U–Pb ages (Figs. 3–5) and geochemical data that test the possibility of earlier extension in the southern GEP. We document a latest Oligocene to Early Miocene phase of extension concurrent with the Early Miocene episode of SMO silicic bimodal volcanism and show that substantial crustal thinning was accomplished before the end of subduction at 14–12.5 Ma. We conclude that distinguishing Neogene magmatism and extension in western Mexico into two stages with a separate subduction and rifting history is incorrect, as crustal extension and decompression-driven mantle melting and crustal melting have been the controlling factors since at least the end of the Oligocene. In this perspective, Basin and Range extension and particularly the bimodal volcanism of the SMO represent the initial stage of a long rifting process that led to the formation of the Gulf of California.

METHODS

Geologic Mapping

Few geological studies exist for the southeastern border of the Gulf of California and, where available, are mostly at a reconnaissance level (Henry and Fredrikson, 1987; Henry, 1989; Ferrari et al., 2002). The first regional geologic synthesis (presented in Ferrari et al., 2007) was
Figure 3 (on this and following page). Histograms and concordia diagrams of U-Pb ages for zircons of silicic rocks. Errors in calculated ages are 2σ. Dashed ellipses are data points not used in calculating the weighted mean. Full details of the U-Pb experiments are given in the Supplemental File (see footnote 1).
based on the available literature data and the integration of 1:250,000 scale maps published by Servicio Geológico Mexicano (SGM) completed in 2002. In the past decade, the SGM has systematically mapped many quadrangles at 1:50,000 scale (available at: http://mapserver.sgm.gob.mx/cartas_impresas/productos/cartas/cartas50/geologia50/numcarta50.html), thus improving the regional geologic database. Although these maps generally have good field control, they often lack absolute age information. For the purpose of this study we have compiled a new regional geologic map through the reinterpretation of the 1:50,000 scale maps of SGM incorporating our new geochronologic data and field work carried out between 2006 and 2010. The map design and integration was accomplished through the use of Google Earth Pro (http://www.google.com) and Quantum GIS (geographic information system; http://www.qgis.org/).

Geochronology and Geochemistry

Previous geochronologic data were limited and unevenly distributed across the studied region. McDowell and Keizer (1977) dated (by K-Ar) ignimbrite successions along the Durango-Mazatlán highway in the central SMO and some of these samples were later redated by 40Ar/39Ar (McDowell and McIntosh, 2012), but without any significant change to the assigned ages. Henry and Fredrikson (1987) and Henry et al. (2003) reported U-Pb (thermal ionization mass spectrometer, TIMS), and K-Ar ages for plutonic and volcanic rocks in southern Sinaloa.

In Ferrari et al. (2002), 40Ar/39Ar ages of the main ignimbrite sequences in northern Nayarit were provided. Additional sparse K-Ar and 40Ar/39Ar biotite and feldspar ages mostly without geologic context were presented in Damon et al. (1979), Solé et al. (2007), and Iriondo et al. (2003, 2004). However, large areas of the SMO in Sinaloa still lack geochronologic data.

The main problem with the K-Ar dating method is its inability to detect thermal resetting or excess Ar (Faure and Mensing 2005). As a further complication, laser ablation–inductively coupled plasma–mass spectrometry (LA-ICP-MS) zircon dating of SMO ignimbrites (Bryan et al., 2008) revealed significant age discrepancies, beyond analytical error, between U-Pb zircon and K-Ar or 40Ar/39Ar biotite and feldspar ages. In several cases, the discrepancy

![Figure 3](https://pubs.geoscienceworld.org/gsa/geosphere/article-pdf/9/5/1161/3343505/1161.pdf)
reflected the incorporation of xenocrystic and antecrystic zircons that skewed the population age toward ages older than the eruption. Inheritance signatures are a particular problem for zircons from Early Miocene rhyolites (see also Ramos-Rosique et al., 2013; Murray et al., 2013), which are suspected to be extensive in the study area. A fundamental conclusion of these recent studies in neighboring areas in the SMO is the importance of stratigraphic control on dated samples and the requirement to double date by both 40Ar/36Ar and LA-ICP-MS techniques to obtain stratigraphically relevant ages (Bryan et al., 2008), an approach that we have followed in this study for critical samples. We are aware that the term antecryst is not unequivocally used in the literature. In this paper we adopt the definition in a previous paper on the SMO silicic volcanism (Bryan et al., 2008), where it was pointed out that when magmatism has been sustained for 15–20 m.y. at a provincial scale, defining what is antecrystic and what is xenocrystic becomes blurred. For ancient magmatic systems like the SMO where individual volcanoes and their erupted products cannot always be identified and correlated, coupled with the lower precision of the LA-ICP-MS technique, it is not possible to discriminate zircons into antecrystic and xenocrystic at time scales <1 m.y., as has been achieved at modern silicic volcanoes (e.g., Charlier et al., 2005; Schmitt et al., 2010). Within the SMO, any zircon younger than 38 Ma has been produced by magmatism associated with the province (considered as antecrysts in this study), while older zircons have been assimilated from mid-lower crustal rocks and are obviously unrelated to the SMO volcanism (xenocrysts).

For this study we collected a large suite of samples from representative rock units exposed along the southeastern margin of the Gulf of California, with the aim of better constraining extensional faulting and volcanism in space and time. Three samples were also obtained from rifted continental blocks submerged in the Gulf of California during the National Science Foundation–funded DANA and ROCA cruises in 2004 and 2008. We obtained 43 new mineral ages by U-Pb LA-ICP-MS and 40Ar/39Ar methods. U-Pb ages were determined at Centro de Geociencias, Universidad Nacional Autónoma de México (UNAM), campus Juriquilla. Ages are summarized in Table 1. Procedures are described in Appendix 1, and complete analytical results are presented in the Supplemental File. The 40Ar/39Ar dating was performed at the Geochronology Laboratory of the Departamento de Geología, Centro de Investigación Científica y de Educación Superior de Ensenada (CICESE) with the procedures described in Appendix 2.

Figure 4 (on this and following two pages). 40Ar/36Ar versus 39Ar/40Ar correlation diagrams for the samples analyzed. First experiment is in pale blue and second experiment is in yellow. One-step laser fusion experiments from sample DANA 46a are in pink. Letters on ellipse identify fraction of gas released reported in the Supplemental File (see footnote 1). The isochron age (t_c) and (40Ar/36Ar)_i calculated is given for each sample. MSWD is the mean square of weighted deviates and n is the number of data fitted. Full details of the 40Ar/36Ar experiments are given in the Supplemental File (see footnote 1).
The \(^{40}\text{Ar}^{39}\text{Ar}\) ages are summarized in Table 1. All the relevant \(^{40}\text{Ar}^{39}\text{Ar}\) information and a discussion of each experiment are given in the Supplemental File (see footnote 1).

A selection of samples dated in this work and in Ferrari et al. (2002) plus other samples for which we have stratigraphic controls were analyzed for major and trace elements. Major elements were analyzed with a Siemens SRS-3000 X-ray fluorescence instrument at Instituto de Geología (UNAM), following the procedures outlined in Lozano-Santa Cruz et al. (1995). Trace element analyses were obtained by ICP-MS using a Thermo Series XII instrument at Centro de Geociencias (UNAM) Querétaro, Mexico. Major and trace element analysis of samples DANA 46a, ROCA 3 J 4, and ROCA 24 J 33 were obtained at the GeoAnalytical Laboratory of Washington State University. Results are presented in Table 2. Further details of procedures and measurements are given in Appendix 3.

**REGIONAL GEOLOGIC SETTING**

The study region encompasses the western part of the southern and central SMO (as defined in Ferrari et al., 2007). The region is here divided into three domains: northern Nayarit (Figs. 6 and 7), southern Sinaloa, and northern Sinaloa (Figs. 8–11), based on the dominant dip and tectonic transport direction during Gulf of California rifting (Fig. 2) (e.g., Axen, 1995) as well as the type of underlying basement. The prevolcanic basement in northern Sinaloa consists of metavolcanic and metasedimentary assemblages of Paleozoic to Late Jurassic age (El Fuerte Group; Mullan, 1978; Vega-Granillo and Vidal-Solano, 2008, 2012; Keppie et al., 2006) covered by shale, marl, and limestone of Berriasian to Turonian age (Servicio Geológico Mexicano, 1999a, 2000). In southern Sinaloa deformed granitoids intruding phyllitic sandstone, quartzite, and quartz-biotite-muscovite schist yielded Jurassic to Early Cretaceous ages (Henry et al., 2003) and are covered by limestone of Alban–Cenomanian age (Bonneau, 1970). In northern Nayarit, prevolcanic basement consists of undated slate and phyllites only exposed in small outcrops 14 km northwest of Acaponeta (Servicio Geológico Mexicano, 1999b) (too small to be shown in Fig. 6). Continental magmatism that formed the rocks of the lower part of the SMO (the lower volcanic complex of McDowell and Keizer, 1977) ranges in age from Cenomanian age (Bonneau, 1970). In northern Nayarit (Figs. 6 and 7), southern Sinaloa, and northern Sinaloa (Figs. 8–11), based on the dominant dip and tectonic transport direction during Gulf of California rifting (Fig. 2) (e.g., Axen, 1995) as well as the type of underlying basement. The prevolcanic basement in northern Sinaloa consists of metavolcanic and metasedimentary assemblages of Paleozoic to Late Jurassic age (El Fuerte Group; Mullan, 1978; Vega-Granillo and Vidal-Solano, 2008, 2012; Keppie et al., 2006) covered by shale, marl, and limestone of Berriasian to Turonian age (Servicio Geológica Mexicana, 1999a, 2000). In southern Sinaloa deformed granitoids intruding phyllitic sandstone, quartzite, and quartz-biotite-muscovite schist yielded Jurassic to Early Cretaceous ages (Henry et al., 2003) and are covered by limestone of Alban–Cenomanian age (Bonneau, 1970). In northern Nayarit, prevolcanic basement consists of undated slate and phyllites only exposed in small outcrops 14 km northwest of Acaponeta (Servicio Geológica Mexicana, 1999b) (too small to be shown in Fig. 6). Continental magmatism that formed the rocks of the lower part of the SMO (the lower volcanic complex of McDowell and Keizer, 1977) ranges in age from Cenomanian age (Bonneau, 1970). In northern Nayarit, prevolcanic basement consists of undated slate and phyllites only exposed in small outcrops 14 km northwest of Acaponeta (Servicio Geológica Mexicana, 1999b) (too small to be shown in Fig. 6). Continental magmatism that formed the rocks of the lower part of the SMO (the lower volcanic complex of McDowell and Keizer, 1977) ranges in age from Cenomanian age (Bonneau, 1970). In northern Nayarit, prevolcanic basement consists of undated slate and phyllites only exposed in small outcrops 14 km northwest of Acaponeta (Servicio Geológica Mexicana, 1999b) (too small to be shown in Fig. 6). Continental magmatism that formed the rocks of the lower part of the SMO (the lower volcanic complex of McDowell and Keizer, 1977) ranges in age from Cenomanian age (Bonneau, 1970).
The northern Nayarit domain is bounded to the south by a left-lateral transpressional zone located just to the north of the Trans-Mexican Volcanic Belt (Ferrari, 1995; Ferrari et al., 2002) and to the north by the Río Mezquital lineament, a fault zone accommodating opposite tilting of strata (Fig. 2). The northern Nayarit domain is a footwall segment in the terminology of Axen (1995), with rocks dominantly dipping east to east-northeast due to north-south– to north-northwest–striking, largely west-dipping normal faults (Figs. 6 and 7). The presence of the lower volcanic complex is evidenced by a small granite body 20 km north of Huajicori (Fig. 6) that yielded a late Maastrichtian U-Pb zircon age (Duque-Trujillo et al., 2013), by the rhyolitic dikes are emplaced along the north-northwest–striking, largely west-dipping normal faults (Figs. 6, 7A, and 7B). The rhyolitic dikes are emplaced in a belt along the western margin of the SMO (ca. 22–18 Ma; Figs. 6 and 7).

Extensional faulting affected almost the entire northern segment apart from a north-elongated elliptical area, ~25 × 35 km wide, centered in the Mesa del Nayar, where ignimbrites are flat lying. At least two generations of high-angle normal faults can be recognized: (1) north-south–striking faults affecting the entire region and bounding grabens and half-grabens (Atengo, Jesús María, Huajicori, Puente de Camotlán, Huajimic, and Sierra de Alica half-grabens; La Ventana and San Agustín grabens); and (2) north-northwest–striking, faulting systems; Fig. 6). Constraints on the age of faulting are available in a number of areas and indicate prolonged extension beginning at the Oligocene-Miocene boundary.

**Atengo Half-Graben**

Faulting in the Atengo half-graben has resulted in tilts to 25° for ignimbrites as young as 28 Ma (Ferrari et al., 2002). On the western side of the half-graben, approximately north-south rhyolitic dikes, 5–10 m thick, cut across the ignimbrite sequence and fed a large rhyolitic dome dated as 27.9 Ma (Ferrari et al., 2002) (Figs. 6, 7A, and 12A). The rhyolitic dikes are emplaced along the flexure separating tilted blocks on the east from flat-lying ignimbrites in the west, so they can be considered to mark the inception of the extension. Several microporphyritic basaltic lavas surround and partly cover the rhyolites. These lavas are only moderately tilted (~10°) and show an intraplate affinity (see Geochemistry discussion).
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<tr>
<th>Sample</th>
<th>Location</th>
<th>Material*</th>
<th>Elevation (m)</th>
<th>Rock type</th>
<th>Material dated</th>
<th>Error (2σ)</th>
<th>Error (2σ)</th>
<th>U-Pb age (mean 206U/238U)</th>
<th>Ar-Ar age</th>
<th>Observations</th>
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<td>Andesitic flows below ignimbrite succession of sample ES 11, near San Pablo, Durango</td>
<td>Basaltic andesite</td>
<td>1948</td>
<td>Plag</td>
<td>33.82</td>
<td>0.28</td>
<td>Isochron age</td>
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<td>ES 11</td>
<td>Pink ignimbrite tilted 35° to the east-northeast by major north-northwest normal fault near San Pablo, Durango</td>
<td>Quartz-rich, strongly welded ignimbrite</td>
<td>1948</td>
<td>Zr</td>
<td>32.5</td>
<td>0.21</td>
<td>MSWD = 0.89, n = 21. Crystals in the range 33.4-31.3 Ma</td>
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<td>SIN 05</td>
<td>Block and ash flow from dome cut by the Concordia graben faults near El Verde, Sinaloa</td>
<td>Glassy rhyolite lava</td>
<td>115</td>
<td>Zr</td>
<td>31.5</td>
<td>0.4</td>
<td>MSWD = 1.4, n = 32. Crystals in the range 36-30 Ma</td>
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<td>SIN 07</td>
<td>Rhyolite dome North of San Juan de Jacobo along a possible northeast-trending fault, Sinaloa</td>
<td>Rhyolite lava</td>
<td>160</td>
<td>Zr</td>
<td>31.5</td>
<td>0.4</td>
<td>MSWD = 1.5, n = 20. Crystals in the range 35-29 Ma</td>
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<td>SIN 18</td>
<td>Las Adjuntas dome along the El Salto–Espinazo del Diablo road, Durango</td>
<td>Flow-banded rhyolite lava</td>
<td>2778</td>
<td>Zr</td>
<td>29.5</td>
<td>0.3</td>
<td>MSWD = 1.3, n = 15. Crystals in the range 30.5–28 Ma</td>
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<td>HUA 1</td>
<td>Ignimbrite from the lower part of the succession along the Huajicori-Picachos road, Nayarit</td>
<td>Crystal-rich ignimbrite</td>
<td>145</td>
<td>Bio</td>
<td>26.48</td>
<td>0.15</td>
<td>Plateau age</td>
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<tr>
<td>MdCH 03</td>
<td>Intermediate lava sequence underlain rhyolite dome of sample MdCH 05, Nayarit</td>
<td>Daolite lava</td>
<td>730</td>
<td>Zr</td>
<td>22.7</td>
<td>0.8</td>
<td>Many discordant zircons. MSWD = 0.31, n = 3</td>
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<tr>
<td>TS 16</td>
<td>Basaltic lava flow underlying 23.5 Ma ignimbrite east of Las Canoas, Zacatecas</td>
<td>Microporphyritic basalt</td>
<td>2523</td>
<td>Plag</td>
<td>24.38</td>
<td>0.75</td>
<td>Plateau age</td>
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<td>RUIZ 7</td>
<td>Lowermost ignimbrite in the hanging wall of San Pedro fault system; road from Estación Ruiz to Mesa del Nayar, Nayarit</td>
<td>Fine-grained, welded and indurated pink ignimbrite with quartz and sandine</td>
<td>185</td>
<td>Zr</td>
<td>23.6</td>
<td>0.2</td>
<td>MSWD = 1.5, n = 15. One 57 Ma xenocryst.</td>
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<tr>
<td>HUA 2</td>
<td>Ignimbrite from the upper part of the succession along the Huajicori-Picachos road, Nayarit</td>
<td>Lithic ignimbrite</td>
<td>1564</td>
<td>Bio</td>
<td>22.99</td>
<td>0.14</td>
<td>Plateau age</td>
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<td>ESC-7</td>
<td>Eroded lava succession along the Escuinapa–Corral de Piedras road, Sinaloa</td>
<td>Basaltic andesite lava</td>
<td>517</td>
<td>Plag</td>
<td>22.39</td>
<td>2.56</td>
<td>Plateau age</td>
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<td>MdCH 10</td>
<td>Nuño de León dome, on top of 21.1 Ma ignimbrite plateau, Nayarit</td>
<td>Daolite lava</td>
<td>1668</td>
<td>Plag, Zr</td>
<td>22.51</td>
<td>0.90</td>
<td>23.7</td>
<td>0.2</td>
<td>Isochron age. U-Pb MSWD = 2.9, n = 16. Possibly antecyclic inheritance. Four xenocrysts (72–63 Ma). MSWD = 1.17, n = 22. Crystals in the range 24.5–21 Ma</td>
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<td>SIN 39</td>
<td>Silicic domes at northwest tip of Sierra El Infierno, Sinaloa</td>
<td>Rhyolite lava</td>
<td>114</td>
<td>Zr</td>
<td>21.6</td>
<td>0.1</td>
<td>Isochron age</td>
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<td>ORO 2</td>
<td>Welded ignimbrite at roadcut west of El Cajon Hydroelectric Plant, Nayarit</td>
<td>Welded Ignimbrite</td>
<td>415</td>
<td>Bio</td>
<td>20.74</td>
<td>0.44</td>
<td>Isochron age</td>
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<tr>
<td>McBarr 01</td>
<td>Rhyolitic dome in the hanging wall of the San Pedro–Acaponeta fault system, east of San Miguel, Nayarit</td>
<td>Rhyolitic dome</td>
<td>170</td>
<td>Zr</td>
<td>20.8</td>
<td>0.2</td>
<td>MSWD = 1.9, n = 25. Crystals in the range 22.5–20 Ma</td>
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<td>McCH 06</td>
<td>Caramota dome, emplaced in the hanging wall of the Huajicori fault system</td>
<td>Rhyolitic dome</td>
<td>171</td>
<td>Plag</td>
<td>20.25</td>
<td>0.64</td>
<td>Isochron age</td>
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<td>HUA 5</td>
<td>Poorly welded ash flow at Rio Acaponeta, northern exit of Huajicori, Nayarit</td>
<td>Ignimbrite</td>
<td>85</td>
<td>Plag</td>
<td>18.98</td>
<td>0.30</td>
<td>Isochron age</td>
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<td>McCH 09</td>
<td>Mafic lava flow along a normal fault near El Río, Nayarit</td>
<td>Olivine basalt lava</td>
<td>1144</td>
<td>Plag</td>
<td>18.32</td>
<td>0.40</td>
<td>Isochron age</td>
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<td>RUIZ 16</td>
<td>Rhyolitic dome capping the Nayarit succession near the inferred caldera rim, west of Mesa del Nayar, Nayarit</td>
<td>Rhyolitic dome</td>
<td>1592</td>
<td>Plag</td>
<td>17.91</td>
<td>0.20</td>
<td>Isochron age</td>
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(continued)
### TABLE 1. SUMMARY OF NEW U-Pb AND Ar-Ar AGES (continued)

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<tr>
<th>Sample</th>
<th>Location</th>
<th>Long (W)</th>
<th>Lat (N)</th>
<th>Elevation (m)</th>
<th>Rock type</th>
<th>Material dated</th>
<th>Material</th>
<th>Error (2σ)</th>
<th>U-Pb age</th>
<th>Error (2σ)</th>
<th>Observations</th>
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<tbody>
<tr>
<td><strong>Sierra Madre Occidental synextensional volcanism (continued)</strong></td>
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<tr>
<td>RUIZ 34b</td>
<td>Rhyolitic dome covering a normal fault at Santa Cruz de Guayabell, road from Estación Ruiz to Mesa del Nayar, Nayarit</td>
<td>104° 48' 4&quot;</td>
<td>22° 3' 32&quot;</td>
<td>846</td>
<td>Rhyolite lava</td>
<td>Bio, Zr</td>
<td>17.57</td>
<td>0.19</td>
<td>18.4</td>
<td>0.3</td>
<td>Isochron age, U-Pb age based on 5 youngest overlapping grains. MSWD = 0.34, n = 5</td>
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<tr>
<td>PER12</td>
<td>Eroded mafic lavas interlayered with sediments Batequitos–El Divisadero road, Sinaloa</td>
<td>107° 42' 36&quot;</td>
<td>25° 18' 31&quot;</td>
<td>206</td>
<td>Basaltic andesite lava</td>
<td>Gms</td>
<td>17.42</td>
<td>0.77</td>
<td>Isochron age</td>
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<tr>
<td>SIN 09</td>
<td>Sierra El Inferno dome complex, north of El Espinal, Sinaloa</td>
<td>106° 52' 58&quot;</td>
<td>24° 11' 48&quot;</td>
<td>155</td>
<td>Rhyolite lava</td>
<td>Zr</td>
<td>18.6</td>
<td>0.3</td>
<td>Concordia age. Single crystals in the 21–16 Ma age range</td>
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<td>MdCH 05</td>
<td>Cuchara dome, emplaced along normal fault, North of Huajicori, Nayarit</td>
<td>105° 15' 40&quot;</td>
<td>22° 46' 37&quot;</td>
<td>739</td>
<td>Rhyolite lava</td>
<td>Gms</td>
<td>17.41</td>
<td>0.07</td>
<td>25.9</td>
<td>0.2</td>
<td>U-Pb age dominated by antecrysts ranging between 28.5 and 23.5 Ma (MSWD 1.7, n = 28). Ar-Ar plateau age of groundmass considered the correct eruption age</td>
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<tr>
<td><strong>Sierra Madre Occidental synextensional volcanism (continued)</strong></td>
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<td>SIN 25</td>
<td>Mafic lava succession near El Saladillo dam, Sinaloa</td>
<td>106° 42' 4&quot;</td>
<td>24° 7' 12&quot;</td>
<td>202</td>
<td>Basaltic lava</td>
<td>WR</td>
<td>14.01</td>
<td>0.23</td>
<td>Isochron age</td>
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<tr>
<td>SIN 21</td>
<td>Dome near El Palmito (unit C of McDowell and Keitzer, 1977), Durango–Mazatlán road, Sinaloa</td>
<td>105° 49' 51&quot;</td>
<td>23° 34' 21&quot;</td>
<td>2005</td>
<td>Rhyolite lava</td>
<td>Zr</td>
<td>13.7</td>
<td>0.3</td>
<td>MSWD = 1.4, n = 16</td>
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<tr>
<td>SIN 15</td>
<td>Basaltic lava flows covering Sierra el Inferno rhyolitic complex, north of Abuya, Sinaloa</td>
<td>107° 0' 4&quot;</td>
<td>24° 16' 8&quot;</td>
<td>154</td>
<td>Microporphyritic basalt</td>
<td>Gms</td>
<td>13.62</td>
<td>0.17</td>
<td>Isochron age</td>
<td></td>
<td></td>
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<tr>
<td><strong>Coastal offshore lavas postdating Sierra Madre Occidental extension</strong></td>
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<tr>
<td>Dana 46a</td>
<td>Sample dredged on Nayarit scarp</td>
<td>106° 37' 48&quot;</td>
<td>22° 26' 24&quot;</td>
<td>–810</td>
<td>Hawaïite diabase</td>
<td>WR</td>
<td>11.96</td>
<td>0.26</td>
<td>Isochron age</td>
<td></td>
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<tr>
<td>Rocca 3J 5</td>
<td>ROV sample at Tamayo dome, offshore Nayarit</td>
<td>107° 17' 54&quot;</td>
<td>22° 58' 34&quot;</td>
<td>–580</td>
<td>Lithic crystal tuff</td>
<td>Feld</td>
<td>11.70</td>
<td>0.07</td>
<td>Plateau age</td>
<td></td>
<td></td>
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<tr>
<td>LM2</td>
<td>El Muelllecito, Sierra Navachiste, Sinaloa</td>
<td>108° 56' 37&quot;</td>
<td>25° 34' 41&quot;</td>
<td>5</td>
<td>Dacite lava</td>
<td>Gms, Pig</td>
<td>11.34</td>
<td>0.79</td>
<td>Isochron age</td>
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<tr>
<td>Rocca 24J 33</td>
<td>ROV sample south of Pescadero transform</td>
<td>108° 56' 17&quot;</td>
<td>24° 2' 28&quot;</td>
<td>–1585</td>
<td>Microdiorite</td>
<td>Hb</td>
<td>11.29</td>
<td>0.37</td>
<td>Isochron age</td>
<td></td>
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<tr>
<td>PER-8</td>
<td>Flat-lying mafic lavas in a quarry west of Quila, Sinaloa</td>
<td>107° 7' 9&quot;</td>
<td>24° 22' 53&quot;</td>
<td>91</td>
<td>Basaltic andesite lava</td>
<td>Gms</td>
<td>10.94</td>
<td>0.23</td>
<td>Weighted mean</td>
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<td>SC 3</td>
<td>Mafic dike cutting Early Miocene granites and aplite dike on Santa Catalina Island, BC</td>
<td>110° 47' 19&quot;</td>
<td>25° 41' 44&quot;</td>
<td>0</td>
<td>Basalt</td>
<td>Plag</td>
<td>10.89</td>
<td>0.52</td>
<td>Weighted mean</td>
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<tr>
<td>PER 7</td>
<td>Lower part of almost flat-lying lava succession at the base of La Piedra de la Madera peak, Sinaloa</td>
<td>107° 20' 51&quot;</td>
<td>24° 42' 53&quot;</td>
<td>190</td>
<td>Basaltic andesite lava</td>
<td>Gms</td>
<td>10.54</td>
<td>0.20</td>
<td>Plateau age</td>
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<td>HUA-6</td>
<td>Flat-lying lava succession at the base of Las Peñas microwave peak, Nayarit</td>
<td>105° 13' 14&quot;</td>
<td>21° 56' 39&quot;</td>
<td>59</td>
<td>Basaltic andesite lava</td>
<td>Gms</td>
<td>10.45</td>
<td>0.15</td>
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<td>PER14</td>
<td>Flat-lying mafic lavas south of Guamuchil, Sinaloa</td>
<td>108° 0' 7&quot;</td>
<td>25° 9' 2&quot;</td>
<td>33</td>
<td>Basaltic andesite lava</td>
<td>Gms</td>
<td>10.33</td>
<td>0.88</td>
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<td>Rocca F6</td>
<td>Lower unit at Farallón de San Ignacio, offshore Topolobampo, Sinaloa</td>
<td>109° 22' 48&quot;</td>
<td>25° 26' 24&quot;</td>
<td>5</td>
<td>Dacite lava</td>
<td>Zr</td>
<td>9.5</td>
<td>0.3</td>
<td>MSWD = 1.9, n = 9</td>
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<td><strong>Pericos Volcanic Field</strong></td>
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<td>PER1</td>
<td>Lava flow near Las Higueras, road to Lopez Mateos reservoir, Sinaloa</td>
<td>107° 31' 12&quot;</td>
<td>25° 2' 24&quot;</td>
<td>226</td>
<td>Hawaïite lava</td>
<td>Gms</td>
<td>0.884</td>
<td>0.097</td>
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<td>PER-4</td>
<td>Lava flow from cone near Aguamitas de Capirato, Sinaloa</td>
<td>107° 32' 24&quot;</td>
<td>25° 6' 36&quot;</td>
<td>425</td>
<td>Hawaïite lava</td>
<td>Gms</td>
<td>0.843</td>
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<td>PER-6</td>
<td>Recent lava flow at arroyo Huicharavito, road to Badiraguato, Sinaloa</td>
<td>107° 37' 48&quot;</td>
<td>25° 15' 0&quot;</td>
<td>193</td>
<td>Hawaïite lava</td>
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<td>0.585</td>
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<td><strong>Choix Volcanic Field</strong></td>
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<td>CHO 4</td>
<td>Young basaltic lava flow along the road to Huiles reservoir, Sinaloa</td>
<td>108° 23' 20&quot;</td>
<td>26° 49' 31&quot;</td>
<td>207</td>
<td>Basaltic lava</td>
<td>Gms</td>
<td>0.138</td>
<td>0.107</td>
<td>One-step age</td>
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**Note:** MSWD—mean square of weighted deviates; ROV—remotely operated vehicle.

*Material abbreviations: Plag—plagioclase; Zr—zircon; Hb—hornblende; Bio—biotite; Gms—groundmass; WR—whole rock; Feld—feldspar; BC—Baja, California.
suggesting that they were emplaced after extension began. A small amount of plagioclase suitable for dating was separated from one of these basaltic lavas (TS 16). The 40Ar-39Ar analyses yielded a consistent plateau age of 24.38 ± 0.75 Ma (Fig. 5; Supplemental File [see footnote 1]), confirming the inception of extension in the Atengo half-graben in the Late Oligocene.

These basalts underlie the Las Canoas ignimbrite succession, dated as 23.5 Ma by K-Ar (Damon et al., 1979) and as 23.3 Ma by 40Ar-39Ar (Ferrari et al., 2002). Similar aged basaltic lavas also occur in the lower part of the ca. 24 Ma El Salto–Espinazo del Diablo ignimbrite sequence, exposed along the Mazatlán-Durango highway 130 km to the northwest (McDowell and Keizer, 1977; age from McDowell and McIntosh, 2012) (Fig. 8), and below the 23.5 Ma Alacrán ignimbrite exposed in the Bolaños graben 90 km to the southeast (Ramos-Rosique, 2013). Other correlative basalts associated with normal faulting are found in the Rodeo half-graben and in the Nazas area (Luhr et al., 2001).

TABLE 2. MAJOR AND TRACE ELEMENT ANALYSIS OF SAMPLES FROM THE SOUTHEASTERN MARGIN OF THE GULF OF CALIFORNIA

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<th>Pleistocene</th>
<th>Sample</th>
<th>CHO 1</th>
<th>CHO 2</th>
<th>CHO 3</th>
<th>CHO 4</th>
<th>PER 1</th>
<th>PER 3</th>
<th>PER 4</th>
<th>PER 5</th>
<th>PER 6</th>
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<td>B</td>
<td>AB</td>
<td>H</td>
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<td>Rock type</td>
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<td>Major elements (wt%)</td>
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<tr>
<td>SiO2</td>
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<td>Al2O3</td>
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<td>16.35</td>
<td>16.59</td>
<td>16.27</td>
<td>15.66</td>
<td>16.27</td>
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<td>16.65</td>
<td>16.70</td>
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<td>8.99</td>
<td>8.99</td>
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<td>1.75</td>
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Trace elements (ppm) | | | | | | | | | | |
| Sc | 24.9 | 26.4 | 18.9 | 26.0 | 20.5 | 20.5 | 18.8 | 20.5 | 19.7 |
| V | 270.2 | 258.9 | 181.6 | 186.4 | 218.2 | 218.2 | 187.3 | 208.1 | 192.7 |
| Cr | 145.0 | 181.3 | 105.3 | 133.3 | 121.8 | 231.0 | 104.5 | 101.5 | 109.9 |
| Ni | 92.0 | 99.2 | 66.5 | 71.6 | 84.9 | 84.9 | 71.4 | 71.4 | 515.7 |
| Rb | 25.6 | 26.2 | 29.6 | 39.9 | 45.5 | 45.5 | 34.9 | 43.1 | 35.8 |
| Sr | 686 | 648 | 573 | 881 | 746 | 746 | 694 | 741 | 684 |
| Y | 29.3 | 29.4 | 29.9 | 35.0 | 31.1 | 31.1 | 29.0 | 31.6 | 30.9 |
| Zr | 225 | 205 | 247 | 360 | 255 | 255 | 242 | 265 | 260 |

Continued.

CIPW normative minerals (wt%) | | | | | | | | | | |
| q | – | – | – | – | – | – | – | – | – |
| c | – | – | – | – | – | – | – | – | – |
| ne | 4.6 | 7.8 | 6.0 | 6.0 | 6.0 | 6.0 | 6.0 | 6.0 | 6.0 |
| di | 18.1 | 16.9 | 11.5 | 13.0 | 16.2 | 16.2 | 14.2 | 11.6 | 11.6 |
| hy | – | – | 5.9 | – | – | – | – | – | – |
| ol | 13.7 | 15.8 | 10.2 | 10.0 | 11.8 | 14.8 | 12.6 | 13.2 | 11.6 |

(continued)
Early extension in the Gulf of California

Jesús María Half-Graben

To the west of the Atengo half-graben, the Jesús María half-graben drops the Las Canoas succession ~1700 m to the west, where it is found tilted as much as 30° (Fig. 7B). The Jesús María fault system also downthrows the Nayar ignimbrite succession ~1300 m. As defined in Ferrari et al. (2002), the Nayar succession consists of several ignimbrite sheets that thicken toward the Mesa del Nayar area, where at least 11 cooling units form a 1-km-thick sequence with 40Ar/39Ar ages ranging between 19.9 ± 0.4 and 21.2 ± 0.3 Ma. Basaltic andesitic lavas dated as 21.3 ± 0.3 Ma found along the Jesús María fault system (Ferrari et al., 2002) (Fig. 7B) suggest that extension associated with half-graben formation was ongoing at the time. The top of the flat-lying Nayar succession reaches a maximum of ~1950 m elevation just west of Mesa del Nayar. The Nayar succession is not exposed in the Jesús María footwall block, where the top of the 23.5 Ma Las Canoas ignimbrite succession is at ~2450 m elevation.

TABLE 2. MAJOR AND TRACE ELEMENT ANALYSIS OF SAMPLES FROM THE SOUTHEASTERN MARGIN OF THE GULF OF CALIFORNIA (continued)

<table>
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<tr>
<th>Sample</th>
<th>DANA 46a</th>
<th>PER 7</th>
<th>PER 8</th>
<th>PER 14</th>
<th>ES 1</th>
<th>LM 1</th>
<th>ROCA 24J 33</th>
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<td>Coastal Sinaloa</td>
<td>Coastal Sinaloa</td>
<td>Coastal Sinaloa</td>
<td>Northeast of Pueblo Nuevo</td>
<td>Sierra Navachiste</td>
<td>Farallón de San Ignacio</td>
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<td>Classification</td>
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<td>BA</td>
<td>BA</td>
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<th>TiO2</th>
<th>Al2O3</th>
<th>Fe2O3</th>
<th>FeO</th>
<th>MnO</th>
<th>MgO</th>
<th>CaO</th>
<th>Na2O</th>
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by guest

on 01 May 2019
These relations indicate that extension along the approximately north-south faults of the Jesús María half-graben must have initiated before the emplacement of the Nayar succession ca. 21 Ma, so the footwall was already a topographic barrier to the east, preventing further eastward outflow of the Nayar ignimbrites. Extension must have continued for a few million years more to tilt and downthrow the Nayar succession east of Mesa del Nayar (Fig. 7B). Further constraints on the age of faulting in the area just west of Mesa del Nayar come from a rhyolitic dome located 25 km to the southwest, which covers a north-south fault without being cut by the fault (Fig. 12B). A sample from this dome yielded 26 U-Pb ages in the range 23–17 Ma (sample RUIZ 34b; Fig. 3; Table 1). A weighted mean including all the analyses would yield an age of 19.02 ± 0.3 Ma, with a mean square of weighted deviates (MSWD) of 3.7, indicating an age spread beyond the obtained analytical error. A further weighted mean, obtained with the 5 concordant and overlapping youngest analyses, yields an age of 18.4 ± 0.3 Ma (MSWD = 0.34, n = 5), which is interpreted as the best estimation for the last phase of zircon crystallization.
Early extension in the Gulf of California

We dated a large rhyolite dome capping the Nayar succession west of Mesa del Nayar, which is unfaulted and has a mostly flat top (i.e., does not appear tilted) (sample RUIZ 16; Table 1). Plagioclase separated from the rhyolite was step-heated and yielded an isochron of 17.91 ± 0.20 Ma (Fig. 4), indistinguishable within error from the biotite age of the RUIZ 34b dome. These geologic relations suggest that by ca. 17.6 Ma extension associated with the north-south faults had essentially terminated.

for this rhyolite dome. We also dated a biotite concentrate from the same sample by Ar-Ar laser step-heating; the resulting isochron yielded an age of 17.57 ± 0.19 Ma (Table 1; Fig. 4), which we consider as the best estimation of the emplacement age of the dome.

We dated a large rhyolite dome capping the Nayar succession west of Mesa del Nayar, which is unfaulted and has a mostly flat top (i.e., does not appear tilted) (sample RUIZ 16; Table 1). Plagioclase separated from the rhyolite was step-heated and yielded an isochron of 17.91 ± 0.20 Ma (Fig. 4), indistinguishable within error from the biotite age of the RUIZ 34b dome. These geologic relations suggest that by ca. 17.6 Ma extension associated with the north-south faults had essentially terminated.

<table>
<thead>
<tr>
<th>Sample</th>
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<th>ESC 2</th>
<th>ESC 3</th>
<th>ESC 8</th>
<th>HUA 5</th>
<th>MicBarr 01</th>
<th>MicBarr 03</th>
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<td>Camino Mazatlán-Miravalle</td>
<td>San Pedro-Acapetla</td>
<td>West of Llano Grande</td>
<td>Road to Rincón Verde</td>
<td>Huajicori-Picachos</td>
<td>San Pedro-Acapetla</td>
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<td>tuff</td>
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Note: Dash indicates not analyzed. Abbreviations: v.f.—volcanic field; A—andesite; AB—alkali basalt; B—basalt; BA—basaltic andesite; D—dacite; H—hawaiite; KT-B—potassic trachybasalt; R—rhyolite; ign.—ignimbrite.

*K-Ar age from Solé et al. (2007).
†Age of the lithologically similar sample R3J-5, collected at the same site.
K-Ar age from Ferrari et al. (2002). Normative minerals: q—quartz; c—corundum; ne—nepheline; di—diopside; hy—hypersthene; ol—olivine.
Huajicori-Picachos Fault System

The Huajicori area is located in northwestern Nayarit (Fig. 6). Here, a system of north-south–striking and west-dipping normal faults creates an ~1800 m escarpment that bounds the SMO ignimbrite plateau toward the coastal plain (Huajicori-Picachos fault system; Fig. 7C). The lowermost ignimbrite exposed by the fault system is a pink, crystal-rich welded ignimbrite with biotite, quartz, feldspar, and plagioclase. We dated a biotite concentrate from this ignimbrite (sample HUA 1) by $^{40}$Ar-$^{39}$Ar, obtaining a plateau age of 26.48 ± 0.15 Ma (Fig. 5; Table 1) from the weighted mean of the last four fractions of the second experiment. We have also dated some zircons from an aphyric dacitic lava in the same succession ~20 km to the north of HUA 1 (sample MdCH 03). We performed a total of 27 zircon analyses; once filtered, these yielded 18 concordant to variably discordant ages. In particular, the 3 youngest analyses yielded concordant and overlapping results, with a $^{206}$Pb/$^{238}$U weighted average age of 22.7 ± 0.8 Ma (MSWD = 0.31, n = 3) (Fig. 3), consistent with the stratigraphic position of the sample. A succession of ~1000 m of ignimbrites covers the Oligocene ignimbrite and Early Miocene dacite. We dated one of the highest ignimbrites near the Los Picachos site, a moderately welded and poorly indurated ignimbrite with biotite and feldspar phenocrysts (HUA 2) that forms a distinctive erosional morphology with sharp peaks. Duplicate experiments by laser step-heating on a biotite concentrate gave a well-defined plateau age of 22.99 ± 0.14 Ma and an isochron age of 23.19 ± 0.27 Ma; we consider the former the best estimate age for the emplacement of this sample because of its smaller error (Table 1). The ignimbrite succession is capped by a large dacitic dome, from which we separated 31 zircons that gave concordant to slightly discordant U-Pb ages in the range ca. 28–23 Ma and 4 xenocrysts with ages of 72–63 Ma (sample MdCH 10; Table 1; Fig. 3). The $^{206}$Pb/$^{238}$U weighted mean of 16 crystals yields an age of 23.7 ± 0.2 Ma (MSWD = 2.9). The stratigraphic position of this dome above the HUA 2 ignimbrite dated as 22.99 Ma, combined with obvious inheritance, indicates that the weighted mean age is biased by an antecrystic population and thus provides no constraint on timing of dome emplacement. For this reason we also dated a plagioclase concentrate from the same sample by the $^{40}$Ar-$^{39}$Ar method. The sample yielded an isochron age of 22.51 ± 0.90 Ma and a plateau age of 22.15 ± 0.93 Ma (Figs. 4 and 5; Table 1), consistent with the stratigraphic relations, and confirming that this dacitic lava is dominated by relatively subtle
antecrystic inheritance, as observed elsewhere in the SMO (e.g., Bryan et al., 2008).

Several other silicic domes are emplaced in the hanging wall of the Huajicori-Picachos fault system. The Cuchara and Caramota domes are particularly interesting because they were emplaced above one of the north-south faults without being obviously cut; 28 zircons with acceptable discordance (i.e., <10% of deviation from concordia) from the Cuchara dome gave a $^{206}\text{Pb}^{238}\text{U}$ weighted mean age of $25.9 \pm 0.2$ Ma (MSWD = 1.7, n = 28) (sample MdCH 05; Table 1). The somewhat high MSWD is indicative of age dispersion and the presence of multiple grain ages or populations, confirmed by $^{40}\text{Ar}^{39}\text{Ar}$ dating. A groundmass concentrate from the same sample was laser step-heated and a well-defined plateau (14 fractions of 19 steps) gave an age of $17.41 \pm 0.07$ Ma (Fig. 5; Table 1). Given the high quality of this age datum, it must be concluded that all zircons in this dome are inherited from previous igneous episodes (i.e., antecrysts) as represented by other dated rocks exposed in proximity to this unit (e.g., ignimbrite HUA1 and dacite lava MdCH 03; Table 1).

For the Caramota dome (sample MdCH 6; Table 1), two $^{40}\text{Ar}^{39}\text{Ar}$ experiments yielded a reproducible age spectra, with a wide plateau age of $20.54 \pm 0.79$ Ma and an isochron age of $20.25 \pm 0.64$ Ma, which we consider to be the best age estimate. We also dated a microporphyritic olivine basaltic lava vented from one of the approximately north-south faults 15 km east of Huajicori (sample MdCH 09; Table 1). A plagioclase concentrate from this lava was laser step-heated, and yielded a plateau age of $18.34 \pm 0.39$ Ma defined by 5 consecutive fractions with 72.06% of the $^{39}\text{Ar}$ released, identical to the isochron age of $18.32 \pm 0.40$ Ma obtained with the combined data of the two experiments. Further age relations are deduced 45 km northwest of Huajicori near the coast, where basaltic andesitic lavas vented from an approximately north-south fault cover, in angular unconformity, ignimbrites tilted to the east (Fig. 6). A plagioclase concentrate from these basaltic andesite lavas (sample ESC 7; Table 1), was dated by the $^{40}\text{Ar}^{39}\text{Ar}$ method by step-
that in most cases they are Miocene in age. We Oligocene. However our new ages indicate
rari et al., 2002). Geologic maps of the SGM
sion as much as 35° to the east-northeast (Fer-
faults that tilt ignimbrites of the Nayar succes-
systems and form a major breakaway zone
bouding the Gulf of California (Figs. 6, 7A,
11A, and 11B), which define a hanging-wall
part of a dome; 27 zircons yielded ages in the
range 23–24 Ma, with an inherited crystal of ca. 57 Ma. The $^{206}\text{Pb}/^{238}\text{U}$ weighted mean
yielded an age of 23.6 ± 0.2 Ma (MSWD = 1.5,
$n = 15$) (Fig. 3). We have also dated zircons
from a porphyritic intrusion cut across by north-
northwest faults 30 km south of Acaponeta, in
the coastal plain (sample MicBarr 01; Table 1).
The rock is silicic, with quartz, feldspar, and
biotite in a glassy matrix, which locally shows
flow banding and probably represents the inner
part of a dome; 27 zircons yielded ages in the
range 20–22.5 Ma, with a $^{206}\text{Pb}/^{238}\text{U}$ weighted
mean of 20.8 ± 0.2 Ma (MSWD = 1.9, $n = 25$). Other
shallower intrusive bodies exposed by the
Pochotitán and San Pedro–Acaponeta fault sys-
tems yielded ages between ca. 20 and 17 Ma
(Ferrari et al., 2002; Duque-Trujillo et al., 2013).
The described relations indicate that approxi-
mately east-northeast extension associated with
heating in a temperature-controlled Ta furnace,
yielding a plateau age of 22.39 ± 2.56 Ma.

In summary, the new ages obtained in the
Huajicori area, as well as the geologic relations,
indicate that a significant part of displacement
along the approximately north-south faults of the
Huajicori-Picachos system occurred before the
emplacement of the ca. 18.98 Ma Huajicori
ignimbrite and was waning by ca. 17.4 Ma,
when the Cuchara dome was emplaced on
one side of the fault escarpment. Basalts dated
as ca. 18.3 Ma vented from the north-south faults
were likely providing the thermal input to
promote small-volume crustal melting and
emplacement of rhylitic domes. This suggests
that extensional faulting in the area began before
the Early Miocene and was likely waning by the
time the mafic lavas were emplaced (youngest
age within error, 19.83 Ma).

**Pochotitán and San Pedro–Acaponeta Fault Systems**

Two major systems of north-northwest–striking
normal faults postdate the north-south fault
systems and form a major breakaway zone
bouding the Gulf of California (Figs. 6, 7A,
7B, and 7D). They consist of a series of normal
faults that tilt ignimbrites of the Nayar succes-
sion as much as 35° to the east-northeast (Ferrari
et al., 2002). Geologic maps of the SGM
interpret some of the rocks cropping out at the
base of the tilted succession as Paleocene to
Oligocene. However our new ages indicate
that in most cases they are Miocene in age. We
dated a tilted ignimbrite near the southern end
of the Pochotitán fault system (sample ORO 2;
Table 1), and the lowestmost ignimbrite in the
hanging wall of the southern part of the San
Pedro fault system (sample RUIZ 7; Table 1).
Sample ORO 2 is from a moderately welded,
indurated lithic ignimbrite with feldspar, quartz,
and biotite, tilted 15° to the east-northeast. A
biotite concentrate from the ignimbrite was laser
step-heated and we conclude that the resulting
isochron age of 20.74 ± 0.44 Ma is the best erup-
tion age estimate (Fig. 4; Table 1). Sample
RUIZ 7 is a pink, well-indurated, moderately
welded ignimbrite with quartz and sanidine, dip-
ing 20° to the east-northeast. A biotite concen-
trate from the ignimbrite yielded 17 acceptable ages
in the range 23–24 Ma, with an inherited crystal of ca. 57 Ma. The $^{206}\text{Pb}/^{238}\text{U}$ weighted mean
yielded an age of 23.6 ± 0.2 Ma (MSWD = 1.5,
$n = 15$) (Fig. 3). We have also dated zircons
from a porphyritic intrusion cut across by north-
northwest faults 30 km south of Acaponeta, in
the coastal plain (sample MicBarr 01; Table 1).
The rock is silicic, with quartz, feldspar, and
biotite in a glassy matrix, which locally shows
flow banding and probably represents the inner
part of a dome; 27 zircons yielded ages in the
range 20–22.5 Ma, with a $^{206}\text{Pb}/^{238}\text{U}$ weighted
mean of 20.8 ± 0.2 Ma (MSWD = 1.9, $n = 25$). Other
shallower intrusive bodies exposed by the
Pochotitán and San Pedro–Acaponeta fault sys-
tems yielded ages between ca. 20 and 17 Ma
(Ferrari et al., 2002; Duque-Trujillo et al., 2013).
The described relations indicate that approxi-
mately east-northeast extension associated with
the Pochotitán and San Pedro–Acaponeta fault
systems began after ca. 21 Ma and that the faults
served as an ascent path for discrete batches of
silicic magmas between 20 and 17 Ma that either
reached the surface as domes or crystallized at
shallow depth. The upper limit of this extensional
episod is constrained by undeformed basaltic
lavas emplaced along the coast. A 28-km-wide
and 35-km-long basaltic plateau is exposed
northwest of Tepic and west of the Pochotitán
fault system (Ferrari et al., 2000a, 2000b; Figs.
6 and 7A). The plateau consists of a succession
of basaltic lavas as much as 300 m thick with an
almost constant dip of 4°–5° toward the north-
west, in striking contrast to the 20°–35° dipping
ignimbrites exposed to the north and to the east.
The uppermost lavas were dated as between 8.9
and 9.9 Ma (Gastil et al., 1979; Righter
et al., 1995), and paleomagnetic studies indi-
cate that they were not affected by significant
tectonic deformation after their emplacement
(Goguitchaichvili et al., 2002). An even clearer
angular unconformity can be observed at Cerro
Peñitas, 5 km west of Estación Ruiz (Figs. 6
and 7B), where flat-lying basaltic andesite lavas
emplaced in the coastal plain surround a block
of ignimbrites tilted 25° to the west-southwest
(Fig. 7B). Laser step-heating experiments on
a whole-rock sample from this basaltic andesite
(sample HUA 6) yielded a plateau age of
10.45 ± 0.15 Ma (Table 1; Fig. 5). In addition,
many ca. 11–10 Ma and north-northwest–striking
mafic dikes are emplaced along and paral-
el to the Pochotitán and San Pedro–Acaponeta
extensional fault systems (Ferrari et al., 2002;
Frey et al., 2007). Most of the dikes are vertical,
and although some of them dip as much as 70°,
they are not sheared, indicating that their incli-
nation is due to the intrusion into preexisting
normal fault planes that were already inactive.
We thus conclude that extension associated with
the north-northwest–striking Pochotitán and
San Pedro–Acaponeta extensional fault systems
must have ended before ca. 10.5 Ma.

**EXTENSIONAL FAULTING AND SYNEXTENSIONAL MAGMATISM IN THE SOUTHERN SINALOA DOMAIN**

**Geologic and Tectonic Setting**

North of the Río Mezquital lineament the
general architecture of the rift is dominated by
west-southwest–tilted ignimbrites and north-
northwest–striking normal faults (Figs. 8,
11A, and 11B), which define a hanging-wall
segment in the Axen (1995) terminology. This
style of faulting continues to the north up to
the Piaxtla River valley, where ignimbrites are
tilted to the east-northeast (Figs. 2 and 8).
Early extension in the Gulf of California

Figure 6. Regional geologic map of northern Nayarit showing the main extensional structures and new and published ages (rounded at the closest decimal). JM—Jesús María, Pic—Picachos, Ca—Caramota dome, Cu—Cuchará dome.

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Figure 7. (A, B, and C) Gulf-orthogonal geologic sections showing rifting style, tilting of strata, and synextensional volcanism for northern Nayarit (locations in Fig. 6). Note variable vertical exaggeration. Color of geologic units same as geologic maps except for ignimbrites. (D) Aerial picture of the Pochotitán fault system (geologic section A). Picture taken at ~20°31'00"N, 104°35'00"W looking north before the construction of the Aguamilpa reservoir. Note 21–20 Ma ignimbrites tilted as much as 35° to the east-northeast. (E) Aerial picture of the western side of the north-south–trending San Agustín graben (geologic section C). Picture taken at ~22°35'40"N, 104°46'00"W looking N35°W.
Figure 8. Regional geologic map of southern and central Sinaloa showing the main extensional structures and new and published ages.

SIG—San Ignacio graben; CG—Conitaca graben; EDD—Espinazo del Diablo; LA—Las Adjuntas dome; MN—Mala Noche; CC—Cerro Cuadrado.
Figure 9. Regional geologic map of Culiacán and Pericos areas (northern Sinaloa) showing selected geologic units, the main extensional structures, and new and published ages.
Figure 10. Regional geologic map of Los Mochis–El Fuerte areas (northern Sinaloa) showing selected geologic units, the main extensional structures, and new and published ages.
A smaller domain of east-northeast–tilted ignimbrites is observed farther inland in the Tayoltita and Mala Noche mining areas (Enríquez and Rivera, 2001; Aranda-Gómez et al., 2003; this work). The Mezquital lineament also marks the southern limit of a basement high, because to the north, the lower volcanic complex and its host rocks are well exposed or at shallower depths than in the northern Nayarit segment (Fig. 8). In Sinaloa, Cretaceous to Paleogene intrusive rocks likely constitute a large composite batholith, the oldest parts of which were deformed during the Laramide orogeny. Henry et al. (2003) dated a syntectonic intrusion as ca. 105 Ma (TIMS U-Pb on zircon). A more recent study using both U-Pb dating of zircons and 40Ar-39Ar dating of muscovite constrains the deformation in the Mazatlán area between ca. 98 and 93 Ma (Cuéllar Cardenas et al., 2012). Post-tectonic intrusions consist of granodiorite to dioritic bodies that thickened the batholith toward the east in Campanian to Eocene time (Figs. 8 and 11). These are covered by ignimbrites and domes of the Oligocene pulse dated west of Durango between 31.7 and 28.5 Ma and by the 24–23.9 Ma Espinazo–El Salto volcanic sequence (Fig. 8) (McDowell and Keizer, 1977;
Figure 12. Geologic and tectonic relations described in the text. (A) Rhyolitic dike feeding a large dome dated as 27.9 Ma (Ar-Ar; Ferrari et al., 2002) in the western part of the Atengo half-graben. Dikes are 10–15 m thick (person in inset for scale). Picture taken at 22°37′51″N, 104°13′29″W, looking north. (B) North-south–striking normal fault covered by a rhyolitic dome dated as 17.6 Ma (sample RUIZ 34b; see Table 1 for coordinates). (C) View of the Baluarte bridge (world’s second-highest bridge). Note the 400-m-deep canyon cut into the massive, flat-lying, basal unit of the Espinazo–El Salto ignimbrite sequence. Picture taken at 23°10′25″N, 106°25′33″W, looking N15W. (D) View looking north of west-dipping Oligocene ignimbrites at Mazatlán lighthouse (Cerro El Crestón). Picture taken at 23°10′25″N, 106°25′33″W, looking N15W. (E) View to the north-northwest of an angular unconformity between flat-lying Espinazo–El Salto ignimbrite sequence (Cerro Cuadrado to left) and Early Oligocene ignimbrites (32.5 Ma, sample ES 11, to right) in the Presidio River canyon. Picture taken at 23°51′00″N, 105°35′30″W. (F) Angular unconformity between slightly tilted El Salto–Espinazo ignimbrite sequence and strongly tilted pre-Miocene ignimbrites in foreground. Picture taken at 23°32′28″N, 105°45′22″W, looking north-northwest from the eastern side of Baluarte River canyon. (G) Angular unconformity (as in E), but north of the Presidio River canyon looking north-northwest from Ejido San Pablo (23°53′32″N, 105°37′22″W). (H) View of the southeastern dome complex of Sierra El Infierno. The base of the Sierra is at 50 m and the highest peak reaches 1400 m above sea level. Picture taken at 24°04′02″N, 106°48′21″W, looking north. (I) North-northwest–striking mafic dikes crosscutting Late Cretaceous granites and aplite dikes at Santa Catalina Island, Baja California Sur. The dike is dated as 10.9 Ma (sample SC 3; see Table 1 for coordinates). (J) Flat-lying mafic lavas along the Sinaloa coast south of Guamuchil dated as 10.3 Ma (sample PER 14; Table 1). Picture taken at 25°12′31″N, 107°59′26″W, looking west. Height of lava pile is 290 m. (K) Subvertical, north-northwest–striking mafic dike cutting a tilted ignimbrite succession near Pericos (Fig. 9). Dike likely fed flat-lying basaltic lavas like those dated as 10.3 Ma nearby (Fig. 12J). (L) Late Pliocene hawaiite sill at Punta Plaxitla. The sill contains xenoliths of spinel lherzolite, granulites, pyroxenites, and megacrysts. Picture taken at 23°38′54″N, 106°48′15″W, looking north-northwest.
McDowell and McIntosh, 2012). Late Miocene mafic dikes and late Pliocene lavas are observed along the coast. Scattered Late Miocene intraplate basalts are also exposed in the eastern part of the SMO in the Río Chico–Canatlán graben (Fig. 8) (Henry and Aranda-Gómez, 2000).

Extensional structures affect the western half of the SMO, in Sinaloa state, as well as a narrow belt on its eastern side, just west of Durango. As in the northern Nayarit segment, two episodes of extension can be recognized: (1) approximately east-west to west-southwest extension responsible for north-northwest-striking faults observed in the Piaxtla, Presidio, and Baluarte river valleys and, possibly, the coastal area; and (2) subsequent west-southwest extension restricted to the coastal area (Pánuco, Concordia, and Villa Unión half-grabens) and a belt east of the SMO (Río Chico–Canatlán graben) (Fig. 7). Timing of faulting can be inferred from geologic relations in a number of areas described in the following.

**Early Extension-Transstension in Westernmost Durango: Tayoltita–Pueblo Nuevo Fault System**

To the northwest of Mazatlán, the Piaxtla and Presidio Rivers cut ~2-km-deep incisions orthogonal to the Gulf of California coast into the SMO igneous succession. The top of the succession is made by the Espinazo–El Salto volcanic sequence, making a plateau at an average 2700 m of elevation. Originally recognized and dated as ca. 23.5 Ma by McDowell and Keizer (1977), the sequence has been recently redated as 23.8–24 Ma (by 40Ar–39Ar; McDowell and McIntosh, 2012). It consists of a thick package of ignimbrites and tuffs (unit A), covered by massive rhyolites (unit B), basaltic lavas (unit C), and three ignimbrite packages (unit D, E, and F, or plateau series), briefly described in Aranda-Gómez et al. (2003). The sequence can reach 1000 m in thickness in the La Ciudad area (Figs. 8 and 11A) and along the Baluarte River (Fig. 12C), but toward the east, in the El Salto area, only the uppermost three units can be recognized, with an approximate thickness of 350 m. East of La Ciudad, the Espinazo–El Salto sequence is essentially flat lying and only affected by minor faulting. However, our reconnaissance field work shows that it clearly filled a topographic trough that resulted from a previous episode of normal faulting. Along the Presidio River valley, between Mala Noche and Cerro Cuadrado (MN and CC in Fig. 8), a thick succession of ignimbrites and andesites is tilted as much as 35° to the east-northeast, making a spectacular angular unconformity with the ca. 24 Ma Espinazo–El Salto succession, which is almost horizontal (Figs. 12E–12G). We dated a sample of an ignimbrite at ~350 m below the top of the tilted sequence near Ejido San Pablo (sample ES 11; Table 1). The 22 dated zircons range between 31.4 and 33.4 Ma. The 206Pb/238U weighted mean yields an age of 32.5 ± 0.2 Ma (MSWD = 0.89, n = 21) (Fig. 5). This ignimbrite could be correlative with the Registro tuff and other ignimbrites dated by McDowell and Keizer (1977) and McDowell and McIntosh (2012) west of Durango City that underlie the El Salto sequence. Andesitic lavas underlying the ignimbrites were also dated (sample ES 10; Table 1). A plagioclase concentrate from these lavas yielded an age of 33.82 ± 0.28 Ma, confirming the Early Oligocene age of this succession.

Tilting of this Oligocene ignimbrite and andesite succession occurs to the west of a north-northwest-striking and west-dipping breakaway fault zone northwest of El Salto that is part of a regional structure, here named the Tayoltita–Pueblo Nuevo fault system (see following). A prominent north-northwest alignment of rhyolitic domes occurs along this fault zone (Fig. 8). Among them is the massive Las Adjuntas dome (LA in Fig. 8), intersected by highway 40 west of El Salto (Figs. 8 and 11A), previously dated as 28.5 Ma by K–Ar (McDowell and Keizer, 1977; age corrected for modern decay constants); 15 zircons separated from this dome yielded a 206Pb/238U weighted mean age of 29.5 ± 0.3 Ma (MSWD = 1.3, n = 15) (sample SIN 18; Table 1) with single ages between 30.5 and 28 Ma; in comparison to the K/Ar age, this may indicate the occurrence of antecrysts. The emplacement of large-volume, viscous rhyolitic domes coupled with the occurrence of inherited zircons is consistent with stratigraphic, structural, and zircon age relationships observed elsewhere in the SMO (e.g., Bryan et al., 2008; Ramos-Rosique, 2013; Murray et al., 2013), and is indicative of an extensional regime allowing the remelting of upper crustal, potentially still partially molten, plutons as well as an easy ascent of these viscous rhyolitic magmas. On this basis, we consider the Las Adjuntas dome to mark the initial extensional activity along the Tayoltita–Pueblo Nuevo fault system.

Toward the north, this pre-Miocene faulting episode is also reported in the Piaxtla River valley, the next valley to the north (Fig. 8). Here, in the Tayoltita mining district, Eocene to Middle Oligocene rocks are tilted as much as 70° to the east-northeast, whereas Early Miocene ignimbrite tilts do not exceed 30° (Enríquez and Rivera, 2001). Microstructural studies established that the first episode formed a negative flower structure defined by north-south–to north-northwest–striking faults with right-lateral extensional kinematics, distributed in a 4-km-wide belt east of Tayoltita (Horner and Enríquez, 1999). This right-lateral transtensional corridor is the northern prolongation of the breakaway fault zone observed in the Presidio River valley (Tayoltita–Pueblo Nuevo fault system). At Tayoltita, transtensional faults displace the mineralized veins dated as between ca. 38 and ca. 32 Ma (Horner and Enríquez, 1999). This constrains this first extensional episode as Late Oligocene, as inferred in the Presidio valley. The second episode of extension must have occurred after the emplacement of the capping ignimbrite sequence, dated as between ca. 24.5 and 20.3 Ma (Enríquez and Rivera, 2001).

**Coastal Belt of Extensional Faults**

Between Espinazo del Diablo (EDD in Fig. 8) and the coast, the ca. 24 Ma volcanic sequence is also faulted and tilted due to a subsequent extensional episode. In a 60-km-wide belt parallel to the coast, volcanic successions are tilted to the west-southwest by northwest-striking and northeast-dipping normal faults. Pre–ca. 24 Ma rocks are tilted as much as 40° whereas the Espinazo–El Salto ignimbrites dip more gently, not exceeding 20° (Fig. 12F). Approximately 12 km southwest of Espinazo del Diablo, two rhyolite domes aligned along strike of a northwest-striking fault are crossed by highway 40 (Fig. 8). The northernmost dome covers part of the Early Miocene succession and has a subhorizontal base. The lava is a coherent porphyritic rhyolite with plagioclase, lesser alkali feldspar, and hornblende that was considered part of the Espinazo–El Salto succession by McDowell and Keizer (1977; their unit B, undated). However, zircons separated from the rhyolites gave 16 acceptable ages in the range 12.7–14.7 Ma. The 206Pb/238U weighted mean yields an age of 13.7 ± 0.3 Ma (sample SIN 21, MSWD = 1.4, n = 16; Table 1). Because the dome does not appear to be faulted, this age can be taken as a minimum age for the activity of the northwest-striking faults in this area.

Near the coast, northwest-striking faults form several asymmetric grabens. The largest of these extensional basins are the Villa Unión and the Concordia half-grabens (Fig. 8). In the eastern part of the Villa Unión half-graben, gravel fills are tilted as much as 24° and unconformably overlie a 22.6 Ma dacite lava that is tilted as much as 40° (Henry and Fredikson, 1987; Aranda-Gómez et al., 2003). The graben fill is intruded by ca. 11 Ma, north-south–to north-northwest–trending mafic dikes dipping 65° to the east, suggesting that extension began sometime in the Middle Miocene and continued for a short time after 11 Ma (Aranda-Gómez et al., 2003). We tried to constrain the age of the Concordia half-graben by dating two rhyolitic domes.
Geologic and Tectonic Setting

The Piaxtla and Elota Rivers mark a tectonic boundary that apparently displaces the prevolcanic basement and the lower volcanic complex batholith in a more inland position (Fig. 8). Whereas in southern Sinaloa these rocks crop out along the coast, to the north of these rivers they are separated from the Gulf of California by a 35–60-km-wide coastal plain. In this region, extensional structures form a 65–85-km-wide belt of northwest-striking normal faults with block tilting both to the southwest and the northeast. They bound a series of grabens and half-grabens cut into the Late Eocene and Oligocene silicic volcanic succession of the lower volcanic complex and their roof pendants, as well as the prevolcanic basement (Figs. 8, 9, 10, 11C, and 11D). These tectonic basins are filled by continental sediments and by concurrent bimodal volcanism, a geologic setting similar to that of Sonora to the north, where intracratonic extensional basins were developed since the Early Miocene (Gans, 1997; McDowell et al., 1997). The timing of extension in northern Sinaloa, however, has remained unknown. Our reconnaissance study allows documenting the age of extension in two areas.

Conitaca Graben and Sierra El Infierno Dome Complex

Sierra El Infierno is a huge silicic complex made of domes, lavas, ignimbrites, and shallow subvolcanic intrusions that rises from low-lying alluvial deposits of the coastal plain. There have been no previous studies of this silicic complex; it covers an ~600 km² area and some peaks reach 1400 m above sea level (Figs. 8 and 12H). This voluminous amount of silicic magma was fed, at least in part, by large north-south–trending silicic dikes, which are clearly seen in aerial imagery and were observed in the field. Lavas are glassy and flow-banded rhyolites have variable crystal content (commonly K-feldspar, quartz, and plagioclase, sometimes biotite and hornblende) and occasional fragments of pre-existing intermediate lavas. The Sierra El Infierno is emplaced in the southwestern part of the Conitaca graben, a tectonic depression bounded toward the east by a series of north-northwest–striking and west-dipping normal faults (CG in Fig. 8). The graben is filled by indurated sandstone and polymictic conglomerate interleaved with and capped by basaltic lavas that abut the silicic rocks of the Sierra El Infierno. Sandstone and conglomerate are crudely stratified and dip 10°–25°, whereas capping basalts are only gently tilted. The time of faulting is clearly established by the ages of the Sierra El Infierno silicic complex and the basaltic lavas intercalated and capping the continental sedimentary succession.

We dated two domes in the southern and western parts of Sierra El Infierno. Sample SIN 39 is a porphyritic to glomeroporphyritic coherent rhyolite with plagioclase, biotite, and alkali feldspar exposed at the northwestern tip of the Sierra. Dated zircons (n = 25) range between 24 and 21 Ma with a 206Pb/238U weighted mean age of 21.6 ± 0.1 Ma (MSWD = 1.17, n = 22) (Table 1; Fig. 3). Sample SIN 09 comes from a crystal-rich rhyolite with plagioclase and biotite, from which we dated 26 zircons in the range 21–16 Ma with a dominant population yielding a 206Pb/238U weighted mean age of 18.6 ± 0.3 Ma (MSWD = 1.00, n = 25) Ma. If these ages are representative of the entire Sierra El Infierno, the silicic volcanism in this area would overlap with that of northern Nayarit, although with a much larger volume.

Previous age of 15.8 Ma was reported in Iriondo et al. (2004) for a basaltic lava in the Conitaca graben, although the geologic context of this sample was not provided. We dated a sample located at 17 km to the southeast near the dam of the El Salto reservoir (Fig. 8) that belongs to the upper part of the basaltic lavas inside the graben. The lava is a fine-grained microporphryritic basalt with plagioclase, clinopyroxene, and olivine (sample SIN 25; Table 1; Fig. 4). The laser step-heating experiment performed on whole rock yielded an isochron age of 14.01 ± 0.23 Ma, which we consider the best estimation for this rock. Sample SIN 15 is also a plagioclase, clinopyroxene, and olivine-phryic basaltic lava that covers rhyolitic lavas in the western part of Sierra El Infierno. Laser step-heating experiments of the groundmass gave an isochron age of 13.62 ± 0.17 Ma for this sample, which within error almost overlaps the age of SIN 25 (Table 1; Fig. 4).

Basaltic lavas also crop out along the coast just southwest of Sierra el Infierno. These lavas are petrographically similar to those exposed farther inland in the Conitaca graben but are generally flat lying. We dated a groundmass concentrate of a basaltic andesite lava from a quarry near the western tip of Sierra el Infierno (sample PER 8; Table 1; Fig. 5). One laser step-heating experiment was performed on a groundmass concentrate from the sample and our preferred age of 10.94 ± 0.23 Ma is taken from the weighted mean of three consecutive fractions with 84.57% of the 39Ar released (see the Supplemental File [see footnote 1] for details). The geologic relations coupled with our new ages constrain extension in the Conitaca–Sierra El Infierno area to the Early to Middle Miocene.

Several other remnants of flat-lying, fissure-fed basaltic lavas are distributed along the coast between Sierra El Infierno and Culiácn (Figs. 8 and 9). The similarity in the mineralogy and composition of the lavas suggests that these outcrops could have been part of a single 60-km-long belt of mafic lavas. We dated a sample from the lower part of a 380-m-thick succession south of Culiácn (Fig. 8; La Pedrera microwave peak). A groundmass concentrate from a basaltic andesite lava (sample PER 7; Table 1; Fig. 5) was laser-step-heated and the two experiments yielded reproducible results with a plateau age of 10.54 ± 0.20 Ma.

Badiraguato Graben

This is a composite tectonic trough bounded by northwest-striking normal faults cut into Late Cretaceous to Paleocene intrusions of the...
lower volcanic complex and their roof pendants (Fig. 9). To the south, the graben is covered by middle Pleistocene basaltic lavas of the Pericos volcanic field (see following). The Badiraguato graben is filled by sandstone and polymictic conglomerate with intercalations of basaltic lavas and some rholitic domes and lavas. The conglomerate is tilted 15°–38° both to the southwest and the northeast (Figs. 9 and 11D). We dated a basaltic andesite lava collected from an eroded succession of lavas between the Badiraguato and Mocorito grabens (sample PER 12; Table 1; Fig. 4). Although the lava is not in direct contact with the conglomerate it is considered equivalent to the basaltic lavas within the graben on a graphic basis. A groundmass concentrate was step-heatred with the temperature-controlled Ta furnace and we take the isochron age of 17.42 ± 0.77 Ma as the best estimate for this sample. Other basaltic lavas are exposed along the coast, ~30 km to the southwest. Similar to those more to the south, these coastal lavas are flat lying. We dated a basaltic andesite (sample PER 14; Table 1; Figs. 4 and 12J) by step heating the temperature-controlled Ta furnace. The age spectrum is somewhat perturbed, but an isochron age of 10.3 ± 0.88 Ma can be confidently calculated. In summary, the ages of the graben fill and the postextensional lavas indicate that extension in this part of Sinaloa is Early to Middle Miocene in age.

**LATE MIOCENE VOLCANISM OF NORTHERN SINALOA AND OFFSHORE AREAS**

Our new age data define a regional pulse of mafic magmatism just after the end of subduction across the entire Gulf of California region. Previously, mafic lavas in the range ca. 11–9 Ma were reported from western Sonora (Mora-Klepeis and McDowell, 2004) and in the Bahía de Los Angeles, Jarujuay, and Santa Rosalía areas in Baja California (Benoit et al., 2002; Calmus et al., 2003; Conly et al., 2005; Pallares et al., 2007). In addition to those described in the previous section, we have recognized additional Late Miocene (ca. 12–10.5 Ma) mafic lavas and dikes in the northernmost part of Sinaloa at Sierra de Navachiste, at the small Farallón de San Ignacio Island southwest of Los Mochis (Fig. 10), within the Gulf of California at Santa Catalina Island, and in submerged rifted blocks (Fig. 2).

Sierra de Navachiste is an erosional remnant of a volcanic field made of basaltic, andesitic, and dacitic lavas and some intermediate domes that currently cover an area of 170 km² along the coast. Episodic deposits produced by the erosion of the lavas are also widespread. The lavas are essentially flat lying and their base is at sea level. We dated a groundmass concentrate from a dacitic lava interbedded with more mafic lavas (sample LM 2; Table 1; Fig. 4). The preferred age is taken from the isochron age of 11.34 ± 0.79 Ma calculated excluding the fractions of suspected excess Ar (see the Supplemental File [see footnote 1]). Similar ages were obtained by the K-Ar method by Pallares et al. (2007) and Gastil et al. (1979) for andesitic lavas in the northwestern part of the field near Topolobampo (Fig. 10).

Farallón de San Ignacio is a 0.3 km² island 25 km offshore Sierra de Navachiste and just north of the northwest fault scarp bounding the Farallón Basin (Figs. 2 and 10). On the island, a massive rholitic lava is covered by 2 basaltic lavas 5–8 m in thickness. The tops of the basaltic lavas are horizontal, and they cover the original irregular morphology of the rholite without any visible paleosol. The rocks are too altered to be dated by ⁴⁰Ar-³⁹Ar, but we separated 11 zircons from the underlying rholitic lava. Single zircon ages of sample ROCA F6 range from 9 to 10 Ma with a ⁵⁷⁴⁰Pb/³⁹U weighted mean age of 9.5 ± 0.3 Ma (MSWD = 1.9, n = 9). Despite this age not being very precise, it nevertheless suggests that the overlying flat-lying basalts may be correlative with those of coastal Sinaloa. In the seismic images provided by Miller and Lizarralde (2013) in the nearby Guaymas Basin, a 2-km-thick evaporite deposit overlies seaward-dipping reflectors. We interpret the seaward-dipping reflectors to correlate with the ca. 12–10 Ma basaltic lavas at Farallón de San Ignacio and along the coast of Sinaloa. Similar basalts, dated as ca. 11–9.1 Ma (Conly et al., 2005), were also found in the Santa Rosalía area (Fig. 2) below the Boleo evaporites that Miller and Lizarralde (2013) correlated with those seen on seismic sections.

Santa Catalina Island is part of a rifted continental block at the latitude of Sierra de Navachiste in the western part of the Gulf of California (Fig. 2). The bulk of the island consists of Late Cretaceous granite of the Peninsular Ranges batholith heavily intruded by Miocene granitoids and aplitic dikes (Pitiro-Lajas, 2009). Many northwest-striking, southwestern-dipping mafic dikes exposed in the cliffs of the northwestern part of the island cut all other rocks (Fig. 12I). A plagioclase concentrate from one mafic dike was laser step-heated and analyzed in two experiments (sample SC 3; Table 1; Figs. 4 and 5); although the sample shows evidence of excess argon, we obtained a plateau age of 10.89 ± 0.52 Ma consistent with the less-precise isochron age of 10.14 ± 2.59 Ma (see the Supplemental File [see footnote 1] for details) that we consider representative of the mafic dike swarm at Santa Catalina Island.

Mafic dikes and silicic rocks of similar ages were also found in the rifted blocks submerged in the Gulf of California. Sample ROCA 3J 5 (Table 1; Figs. 4 and 5) is a lithic tuff that was collected using the remotely operated vehicle (ROV) Jason at the top of the Tamayo dome (also called Tamayo Bank; Sutherland et al., 2012), a 30 × 40-km-wide, ~1000-m-high volcanic structure 80 km offshore Mazatlán (Fig. 2). Two laser step-heating experiments were performed on a feldspar concentrate defining a plaeau age of 11.70 ± 0.07 Ma indistinguishable from the isochron age obtained by combining all the data points from the two experiments. Silicic volcanism of Late Miocene age is widespread in Sonora (Mora Alvarez and McDowell, 2000; Vidal-Solano et al., 2005, 2007), but has never been reported to the south. The sample from the Tamayo dome is the first evidence of Late Miocene silicic pyroclastic rocks in the southern part of the Gulf of California. Sample ROCA 243-33 was also collected using the ROV on a rifted continental block cut by a transform fault bounding the Pescadero Basin to the south (Fig. 2). It is a porphyritic, fine-grained subvolcanic microdiorite with plagioclase and hornblende (0.25–3.5 mm), scarce biotite and opaque minerals, probably representing a shallow sill. A hornblende concentrate was laser step-heated and we consider the isochron age of 11.29 ± 0.37 Ma as the best estimation for this sample.

Sample DANA 46a was dredged from a prominent north-northeast–striking and 800-m-high fault scarp offshore northern Nayarit, ~90 km southeast of the Tamayo dome (Fig. 2). The sample is a mafic aphyric lava (either a flow or a dike) presumably associated with the fault scarp. Four experiments with different mass spectrometers and heating methods gave very consistent results, yielding a 11.96 ± 0.26 Ma isochron age (Table 1; Fig. 4; Supplemental File [see footnote 1]). This age overlaps with those obtained for northwest-striking mafic dikes in southern Sinaloa and northern Nayarit (Henry and Aranda-Gómez, 2000; Frey et al., 2007). In seismic profiles, Sutherland et al. (2012) recognized an irregular, ropey layer mantling the seismic basement offshore of northern Nayarit that they interpreted as a volcanic layer of the Comodú Group (24–12 Ma). Our dated samples ROCA 3J 5 and DANA 46a, collected close to the trace of the seismic profile, provide a direct constraint of the nature and age of the uppermost part of this volcanic layer.

**LATE PLIOCENE AND PLEISTOCENE INTRAPLATE VOLCANISM**

Onshore post–Late Miocene volcanism is rare in the southeastern Gulf of California. The only previously known occurrences were the 3.4 Ma Mesa Cacaxtla shield volcano and...
the 2.1 Ma mafic lavas at Punta Piaxtla (Aranda-Gómez et al., 2003), both located along the coast ~60 km northwest of Mazatlán (Fig. 8). The geologic maps of the SGM also report two other mafic volcanic fields with young morphologic appearance, Pericos and Choix. The only previous age available was a K-Ar age of 0.7 Ma (Clark, 1976) for a lava in the Choix volcanic field, but no geochronologic information was available for the Pericos volcanic field and there was no geochemical study for either field.

The Pericos volcanic field covers an area of ~20 x 32 km, and is ~25 km north of Culiacán (Fig. 9). Lava flows, small shield volcanoes, and cinder cones of basaltic composition unconformably overlie Paleocene granodiorite and Oligocene ignimbrites, which are mostly tilted to the west-southwest. Lavas are porphyritic and contain olivine, plagioclase, and clinopyroxene in a microcrystalline matrix. Some lavas contain abundant megacrysts of green clinopyroxene (to 8 cm), olivine (to 1 cm), and/or plagioclase (to 1 cm), or aggregates of olivine and clinopyroxene. We recognized at least 16 cinder cones and maars, all with a well-preserved morphology, suggesting a Pleistocene age for this volcanism. We dated three samples, trying to cover the entire age range of the field. The samples (PER1, PER4, and PER 6) are hawaiitic in composition and have low K contents, so Ar had to be released in a few steps. Nevertheless, they provide consistent middle Pleistocene ages ranging between 884 ± 97 and 585 ± 82 ka (Figs. 4 and 5; Table 1, see the Supplemental File [see footnote 1] for details). Rocks from the late Pliocene Punta Piaxtla sills and Mesa Cacaxtla shield volcano, 200 km to the south-southeast along the Sinaloa coast, show remarkably similar compositions, mineralogies, and megacryst contents (see following), suggesting that the mantle of this wide region achieved a homogeneous composition that has been maintained during the past 3.4 m.y.

The Choix volcanic field consists of a few cinder cones and lava flows 120 km north-northeast of Los Mochis in the northernmost corner of Sinaloa, near the boundary with Sonora and Chihuahua (Fig. 9). Lavas are emplaced on top of Paleocene granodiorite and ignimbrites that are undated, but probably Oligocene. Sample CHO 4 comes from an alkali-basalt lava from a very well preserved cinder cone near the Huites reservoir dam (Fig. 9). The lava is vesiculated with phenocrysts of plagioclase, olivine, and pyroxene, and megacrysts of pyroxene (to 3 cm) and olivine (to 0.5 cm). Due to the very low K content and the presumed young age, only four fractions were collected. The bulk of the ³⁹Ar was released in the last two fractions and no plateau can be defined. This also yielded an imprecise isochron age of 29 ± 181 ka. The best estimate for this sample is 138 ± 107 ka, taken from the third fraction containing the bulk of the ³⁹Ar released (see the Supplemental File [see footnote 1] for further details). Although imprecise, this age confirms the very young age of the Choix volcanic field.

GEOCHEMISTRY

Major and trace element analyses were obtained for volcanic rocks of latest Oligocene to Early Miocene, Late Miocene, and Pleistocene age (Table 2) distributed along the southeast margin of the Gulf of California.

Late Oligocene to Early Miocene (28–16 Ma)

Late Oligocene to Early Miocene samples display a bimodal distribution in the total alkali versus silica diagram (Fig. 13A; LeBas et al., 1986). Samples with low silica content (SiO₂ = 48.7–55.2 wt%) are lavas and dikes that are classified as hypersthene-normative basalts and quartz-normative basaltic andesite, whereas silicic samples (SiO₂ = 68.2–83.0 wt%) from domes, lavas, and ignimbrites are classified as peraluminous (molar Al₂O₃/CaO + Na₂O + K₂O, A/CKN >1), corundum-normative dacite and rhyolite, with a predominance of rhyolitic compositions. Most samples plot in the field of subalkaline rocks of Irvine and Baragar (1971), with the exception of one basaltic lava from the Atengo half-graben (TS 16), which has the lowest SiO₂ content and plots in the field of alkaline rocks (Fig. 13A).

None of the mafic samples can be regarded as primitive mantle melts, as they have relatively low MgO and Cr contents (Figs. 13B, 13D) and low Mg# (100 × molar MgO/MgO + FeO total) of 42.1–52.5 (Table 2). The TiO₂ content is variable (Fig. 13C), and is highest for the alkaline sample (TS 16; 2.39 wt%).

Mafic rocks have variable trace element patterns in multielement diagrams normalized to primitive mantle values (Sun and McDonough, 1989) (Fig. 14E). Negative anomalies in Nb and Ta are absent in the TiO₂-rich alkaline sample (basalt TS 16), and are variably developed in the subalkaline basalts and basaltic andesite. Positive Pb and Sr anomalies are present in all samples, but are more pronounced in the basaltic andesite. The lack of a Nb-Ta anomaly in sample TS 16 indicates an origin in an intraplate setting; nevertheless, the enrichment in Pb and Sr, which is not a common feature of intraplate magmas, could reflect assimilation of crustal material (feldspar contamination) by the mafic magma. Trace element patterns of the subalkaline samples have typical features of subduction-related rocks; however, in this sample group, crustal assimilation processes can have also contributed to generate signatures similar to those found in arc volcanic rocks. Alternatively, the subduction signature can be inherited from mantle domains previously modified by subduction components, which partially melted under the extensional regime that generated the intraplate magmas. Consequently, the subduction-related geochemical signatures are unlikely to provide unequivocal constraints on tectonic setting.

Silicic rocks are enriched in the most incompatible elements (Rb-U) and display well-developed negative Nb-Ta and positive Pb anomalies (Fig. 14F). Despite the high silica content of these rocks, negative Ba anomalies, which result from the fractionation of K-feldspar, are absent. Eu and Sr negative anomalies are small in dacitic rocks, implying negligible plagioclase fractionation. The trace element behavior indicates that crystal fractionation had a restricted role in the early evolution of the silicic magmas; this is supported by the fact that inherited zircons remained in the melt. This feature is often associated with crustally derived magmas (e.g., Miller et al., 2003; Bryan et al., 2008). An origin in the crust is also consistent with the bimodal composition of the magmas emplaced in this period.

It is important to note that intermediate rocks comparable with the middle and upper members of the Comondú Group (19–12 Ma) observed in Baja California were not observed along the eastern margins of the Gulf of California through Nayarit and Sinaloa. In this region volcanism is essentially bimodal, represented by rhyolitic domes and surrounding basalts in the range ca. 21–13.5 Ma and thus overlapping with peak activity of intermediate composition volcanism of the Comondú Group. This supports the observation of Bryan et al. (2013) that, regionally, bimodal volcanism persisted during the Middle Miocene and to the east of the lopus of Comondú Group volcanism.

Late Miocene (11.6–9 Ma)

Volcanic rocks emplaced in the Late Miocene have mainly a bimodal compositional distribution (Fig. 13A). Mafic to intermediate samples (SiO₂ = 51.4–57.2 wt%) are classified as subalkaline quartz-normative basaltic andesite, low-silica andesite, and potassic trachybasalt. Unlike rocks emplaced in the previous event, basalts are absent and more differentiated compositions with lower MgO and Cr contents (Figs. 13B, 13D) and Mg# (36.7–53.5) occur. Analyses of scarce rocks with SiO₂ > 62 wt% indicate metaluminous (A/CKN < 1) high-silica andesite, dacite, and rhyolite compositions.
Low-silica rocks from the coastal area and Farallón de San Ignacio display relatively similar, subparallel trace element patterns in a mantle normalized multielement diagram (Fig. 14B), with weakly developed Nb-Ta anomalies, positive anomalies of Ba, Pb, and Sr, and relatively flat rare earth element (REE) patterns (chondrite normalized La/Yb_n = 3.7–5.3). Very similar compositions are reported for the nearest known outcrops of similar age to the north of the study area, in the central Sonora coast (Till et al., 2009), and for the Santa Rosalía area in Baja California (Conly et al., 2005) (Figs. 1 and 14B). Regional compositional variations are suggested by one potassic trachybasalt from the submarine Nayarit scarp in the southeastern Gulf of California (Fig. 2), which has the lowest SiO_2 content (51.4 wt%) and is more enriched in most trace elements, resulting in a more pronounced Nb-Ta anomaly. In addition, a basaltic andesite collected northeast of Pueblo Nuevo (Fig. 8), within the SMO, has an Nb-Ta anomaly similar to that of the coastal lavas, but is more enriched in the most incompatible elements (Rb to Sr), and has a stronger depletion in the heavy REEs with respect to the light REEs (La/Yb_n = 11.5). A more marked difference is observed in Late Miocene mafic lavas from the Río Chico–Canatlan graben, located in the eastern part of the SMO (Fig. 8) (Henry and Aranda-Gómez, 2000); the lavas have the high Nb contents of intraplate rocks. Differences in composition from east to west indicate changes in the mantle source composition from the Gulf of California to the eastern SMO, probably related to decreasing contributions of subduction components to the mantle and to increasing magma segregation depth.

Late Miocene silicic rocks from coastal and offshore areas display well-developed negative Nb-Ta and positive Pb anomalies, but differ in the abundance of trace elements (Figs. 14C, 14D). A dacitic tuff (R3J-4a) from the submarine Tamayo dome (Fig. 2) has a higher abundance of most elements and displays pronounced negative anomalies in Ba, Sr, and Eu that indicate extensive feldspar fractionation. Similar differences are observed in Middle to Late Miocene silicic rocks (SiO_2 >70 wt%) from the central Sonora coast (Till et al., 2009), and from the Santa Rosalía area in Baja California (Conly et al., 2005) (Figs. 2 and 14D). A markedly different trace element pattern characterizes a Sierra Navachiste dacite sample (Figs. 10 and 14C), which has a much lower Nb-Ta abundance, positive Ba and Sr anomalies, no Eu anomaly, and stronger depletion toward the heavy REEs (La/Yb_n = 13.3). The composition of this sample is similar to those of samples from Sierra Libre, in the central Sonora coast (Till et al., 2009), and also coincides relatively well with that of clinopyroxene-bearing andesite from Santa Rosalía (Conly et al., 2005), and Mg-andesite reported for the Borja and Jaraguay volcanic fields (Pal-lares et al., 2008) in the central Baja California peninsula. The variable composition of the Late...
Miocene silicic rocks point to variable magma generation processes acting during this event. These processes may have involved crustal melting and assimilation, or differentiation of mafic rocks through crystal fractionation, but further studies are needed to better constrain the origin of the observed differences.

Middle to Late Pleistocene (Younger Than 1 Ma)

Lava samples from the Pleistocene Choix and Pericos volcanic fields (Figs. 9 and 10) plot in the field of alkali rocks (Fig. 13A). One sample classifies as hypersthene-normative basalt, and the rest are nepheline-normative hawaiite and alkali basalt. The samples have the highest MgO, TiO$_2$, and Cr contents of all studied samples (Figs. 13B–13D), and Mg$^+$ of 48.2–57.9, but these values are lower than those expected for primary magmas derived from the mantle, implying that differentiation occurred during their ascent.
The samples have quite uniform trace element patterns (Fig. 14A), which resemble those of oceanic island basalts (OIB), with the highest normalized abundances in Nb and Ta and strong enrichment of the light REEs with respect to the heavy REEs (La/Yb $\approx$ 9.4–13.5). The trace element pattern of a late Pliocene hawaiite from Mesa Cacaxtla (Figs. 8 and 14A), reported by Aranda-Gómez et al. (1997), is also remarkably similar to that of Choix and Pericos, considering the different emplacement times (Cacaxtla at 3.2 Ma; Aranda-Gómez et al., 1997; Pericos at 0.585–0.884 Ma; Choix at 0.138 Ma, this work) and the distance between them.

The alkaline character and the trace element composition suggest that the Pleistocene magmas originated in a deeper, enriched OIB-like mantle source in the stability field of garnet, and differentiated at crustal levels.

**DISCUSSION**

**Timing of Extension in the Southern Gulf Extensional Province**

Figure 15 summarizes the ages of extension documented in this work and the literature. A precise determination of the direction of extension awaits a structural study of the kinematics of the various fault systems that is in progress and will be published later. Although a strike-slip component of motion has been observed along some of the faults, the large vertical displacement and the consistent tilting of blocks indicate a dominant normal motion and an almost fault-orthogonal direction of extension. At a regional scale, we recognize a fairly consistent pattern along the ~700 km of length of the southeastern to central GEP with episodes of extension since ca. 29 Ma.

A Late Oligocene (ca. 29–24 Ma) phase of extension is documented in the central part of the study region (southern Sinaloa domain) in the area bounded by the Tayoltita–Pueblo Nuevo fault system to the east and the Concordia fault to the west (Fig. 8). This early extension was probably younger than that reported in the Rodeo and Nazas area (Lurh et al., 2001) and in the northern part of the Rio Chico–Canatlan graben (Loza-Aguirre et al., 2012) to the east of the SMO unextended core (Fig. 15). We expect a similar Late Oligocene extensional phase to have affected areas to the north (northern Sinaloa domain, Fig. 9), particularly in the more inland part of the GEP, which we did not study directly. Supporting this is that extension of this age has recently been documented further to the north in the Guazapares area of southernmost Chihuahua (Murray et al., 2013). Because of the extensive cover of Early Miocene ignimbrites, the existence of a Late Oligocene phase of extension is difficult to prove in the northern Nayarit domain. However, Late Oligocene normal faulting is documented at least in the central Bolaños graben (Fig. 2), where some of the faults (e.g., the prominent Ballena and Cabreras faults) are also northwest trending and served as mineralization pathways during the Early Miocene (Ramos-Rosique, 2013). The direction of extension during this first phase is not well constrained but, considering the right-lateral component of motion reported in the northern part of the Tayoltita–Pueblo Nuevo fault system (Hornor and Enríquez, 1999), it may be slightly oblique with respect to the main faults, i.e., west-northwest–east-southeast directed (Fig. 15).

Extension continued during the Early and Middle Miocene (ca. 24–12 Ma) in Sinaloa and northern Nayarit. In Nayarit, this phase can be further divided into two main episodes with different directions of extension. Normal faulting with approximately north-south trends affected a 180-km-wide area between Bolaños and Huajicori (Fig. 15) between ca. 24 and 18 Ma and was postdated by north-northwest–trending fault systems (Pochotitán and San Pedro–Acaponeta) active between before 18 to 11 Ma along the coast (Fig. 6). In Sinaloa, Early to Middle Miocene extensional faults strike essentially north-northwest, forming grabens and half-grabens filled by 21–17 Ma dome complexes and volcanioclastic sediments interleaved with ca. 17–13 Ma basalts. This last phase of extension is postdated by flat-lying basaltic lavas fed by north-northwest–trending dikes dated as between 11.3 and 10.3 Ma. These geologic and geochronologic relations make the Sinaloa area the southern prolongation of the latest Oligocene–Middle Miocene extensional belt of Sonora, where high-angle faulting and volcanioclastic sedimentation of the Bauca riot Formation occurred from 24 Ma and was essentially over by ca. 12 Ma (McDowell et al., 1997).

The western limit of this early extension is poorly constrained due to the superposed subsequent faulting in the southern Gulf of California. Textural and geologic evidence indicate that Early to Middle Miocene (ca. 21–15 Ma) plutons exposed along the Nayarit coast, the islands offshore southern Baja California, and recovered in the rifted blocks inside the southern Gulf of California were emplaced at shallow depths (Duque-Trujillo et al., 2012). The short time span between U-Pb zircon ages and hornblende and biotite ages also suggest rapid cooling and the possibility that they were emplaced into an extending crust (Duque-Trujillo et al., 2012), but more thermochronologic studies are needed to better constrain this inference. However, Mark et al. (2012) found apatite U-Th(He) exhumation ages of 25–17 Ma in the Loreto area that are in agreement with the early extension of the southern GEP evident from our study. In southernmost Baja California, the Cretaceous batholithic rocks of the eastern part of the Los Cabos block show a geochronologic, geochemical, and isotope similarity with those of the Jalisco block in mainland Mexico (Schaaf et al., 2000) and those exposed in the Maria Madre Island in between (Pompa-Mera et al., 2013), suggesting that the two blocks were contiguous in a north-northwest alignment prior to rifting (Fig. 15). The normal faults of the Los Cabos block are mostly north-south striking, thus parallel to the Early and Middle Miocene faults of Nayarit and Jalisco described in this work. The onset of faulting is constrained by the oldest sediments in the Los Cabos basin (Fig. 15) that consist of red terrestrial sandstone and conglomerate (Calera Formation; Martínez-Gutiérrez and Sethi, 1997). These sediments are not dated, but they overlie tilted silicic volcanic rocks corelated with the ca. 21–19 Ma ignimbrites of the La Paz area. They are also conformably covered by the marine Trinidad Formation, assigned to the Late Miocene on a palaeontological basis (Martínez-Gutiérrez and Sethi, 1997). Based on these stratigraphic constraints, the approximately north-south normal faults in the Los Cabos blocks might be partly coeval with the last extensional faulting in the northern Nayarit domain (ca. 18–12 Ma). However, thermochronologic studies indicate that rapid cooling of the footwall of the Los Cabos fault initiated at the beginning of the Late Miocene (Fletcher et al., 2000; Lovera et al., 2010). In summary, despite the fact that the onset of extension cannot be precisely determined, we consider that the available data do not preclude the possibility that by Middle Miocene time, the area now occupied by southern Gulf of California was already an extending basin.

**Pre–Late Miocene Extension in the Northern Gulf Extensional Province**

The early extension we have documented in this work cannot be limited to the southern-central GEP. Early to Middle Miocene extension is well documented in eastern and central Sonora (McDowell et al., 1997; Gans, 1997; Gonzalez-León et al., 2000; Vega-Granillo and Calmus, 2003; Nourse et al., 1994; Wong and Gans, 2003; Wong et al., 2010), and constitutes the obvious prolongation to the north of the extensional belt of Nayarit and Sinaloa described in this work. However, the western boundary of this extensional belt is uncertain. Micro-paleontological studies of several deep wells drilled in the Wagner, Consag, and Tiburón...
Basins (Fig. 1) reported the occurrence of
>1 km of marine sediments with shallow-water
(<200 m) foraminifera, dinoflagellates, and
nanofossils older than 11.2 Ma (Helenes et al.,
2009). Recent reinterpretation of these data by
J. Helenes and A. Carreño (2013, written com-
mun.) indicates that part of the recovered well
material is reworked, a feature already reported
for other sites in the northern Gulf of Califor-
nia (e.g., Imperial Formation, McDougall et al.,
1999; Laguna Salada, Martín-Barajas, 2001;
Isla Tiburón, Gastil et al., 1999). Given that
the Peninsular Range Batholith of Baja Cali-
fornia was uplifted in the Early Eocene (e.g.,
Axen et al., 2000) the widespread occurrence of
reworked Middle Miocene marine microfossils
requires the existence and proximity of shallow-
water environments within the northern Gulf
of California by the Middle Miocene. This inter-
pretation is apparently at odds with the geology
of Isla Tiburón (Fig. 1) (Oskin and Stock, 2003;
Bennett et al., 2012), where marine incursion
has been well dated as latest Miocene. However,
we note that Isla Tiburón is at the southeastern
margin of a basin that, in its central part, has
>7 km of sediments overlying the Late Creta-
ceous basement (Martín-Barajas et al., 2013).
The well T-1 drilled by PEMEX near the deepest
part of the Tiburón Basin encountered ~4.8 km
of marine deposits (Helenes et al., 2009), which
in their lower part are considered not reworked
and have an upper age limit of 11.9 Ma based
on the microfossils Ciliacargolithus floridanus
(known range 37–11.9 Ma), Cribroperdinium
tenabulatum (69.9–11.63 Ma), Paleocy-
sidotinum golzwense (56–9.2 Ma), and Dap-
silidinium pseudocolligerum (56–7.12 Ma)
(J. Helenes, 2013, written commun.). The recent
seismic profiles of Martín-Barajas et al. (2013)
show that the lowermost sedimentary package
pinches out to the south and is not deposited on
the basement high at the southeastern margin of
the basin (Tiburón shelf and Isla Tiburón).
Therefore, the latest Miocene age of marine
incursion at Isla Tiburón does not contradict
the pre–11.9 Ma marine sediments in the deepest
part of Tiburón Basin and simply records the
widening development of marine conditions
in the gulf in the Late Miocene. A definitive
answer on the existence and extent of localized
marine basins of pre–Late Miocene age in the
northern Gulf of California awaits further stud-
ies of well stratigraphy and seismic data from
the deeper part of the Tiburón and other basins,
but the possibility of an early extension in this
part of the gulf cannot be ruled out.

Pre–Late Miocene Lithospheric Thinning
and Motion of Baja California

The data reported in this work provide impor-
tant constraints on the magnitude of pre–Late
Miocene rifting in the southern GEP. Different
seismic methods consistently indicate that the
clast along the Nayarit and Sinaloa coast is sig-
nificantly thinner than in the core of the SMO.
Estimations of the Moho depth from receiver
functions and seismic refraction along the
coast range between ~21 km west of Culiacán
to ~18 km in northern Nayarit (Persaud et al.,
2007; Lizzarralde et al., 2007; Savage and Wang,
2012) (Fig. 2). By contrast, the crustal thickness
in the unextended core of the SMO is estimated
to be 55 km northeast of Culiacán (Bonner and
Herrin, 1999), ~40 km northeast of Mazatlán.
The recognition of this early phase of extension in the GEP has important implications for the mode of extension and lithospheric rupture that led to the formation of the Gulf of California. Although the kinematics of opening are debated, studies in the past 20 yr essentially assumed a ca. 14–12.5 Ma initiation of the rifting process (e.g., Stock and Hodges, 1989; Gans, 1997; Umhoefer et al., 2001; Oskin et al., 2001; Fletcher et al., 2007; Sutherland et al., 2012), implying a very fast rate of crustal thinning and the initiation of seafloor spreading only ~6–10 m.y. after the onset of extension (Umhoefer, 2011). This is a very short time span for complete rupturing of the lithosphere, which is commonly accomplished over 25–30 m.y. in other examples (see review in Umhoefer, 2011). Our results indicate a more reasonable time span of ~25 m.y. between the initiation of extension and the initial formation of oceanic crust.

An early phase of extension across the GEP beginning ca. 29 Ma also reconciles the apparent discrepancy between the 275–300 km of offset across the northern Gulf of California since 12.5 Ma estimated by correlative geologic units (Gastil et al., 1973, 1991; Oskin et al., 2001; Miller and Lizarralde, 2013) and the ~450–500 km of offset needed to close the southern Gulf of California based on palinspastic reconstructions using an Airy isostatic model of crustal thickness (Fletcher et al., 2003). In fact, the total offset estimated by Fletcher et al. (2003) can be accounted for by the sum of the post–12.5 Ma rifting plus the early extension documented in this work to have initiated ca. 29 Ma. We have tried to estimate the amount of this early extension by looking at the motion of Baja California during this period (Fig. 15). The position of Baja California before extension at 30 Ma is chosen in a way that realigns the Late Cretaceous batholiths of Puerto Vallarta (Jalisco block), Los Cabos, and Sinaloa, as well as by bringing the paleotrench west of Baja California in line with the present trench west of Puerto Vallarta (Fig. 15). In this reconstruction, the southern tip of Baja California is 475 km south-east of its present position, essentially eliminating all the crustal stretching estimated by Fletcher et al. (2003) at the mouth of the Gulf of California. The position of Baja California at 12 Ma is obtained by removing the 245 km of oceanic crust accreted at the East Pacific Rise (Lizarralde et al., 2007) plus 110 km of extension on both sides of the oceanic crust: 35 km in the stretched crust southeast of the Los Cabos block (Páramo et al., 2008) and 75 km of extension in the 200-km-long area of stretched crust between Puerto Vallarta and the continental slope west of Tres Marías Islands (Lizarralde et al., 2007) (Fig. 2). In this scenario, the motion of Baja California between ca. 30 and 12 Ma would amount to 135 km, which, based on the age and geometry of faults, we propose was accommodated by west-northwest to east-west extension between ca. 30 and ca. 18 Ma and by west-southwest extension between ca. 18 and 12 Ma. Our reconstruction implies a moderate rate of separation of Baja California of 7.7 mm/yr during the first phase and 8.3 mm/yr in the second. Of note is that these values are consistent with the rate of extension of the East African Rift (7 mm/yr; Fernandes et al., 2004). The subsequent 355 km of northwestern motion would have been accomplished with a higher average rate of 29.5 mm/yr. However, if we consider the spreading rate for the past 3.6 m.y. in the southern Gulf of California (45.1 ± 0.8 to 51.1 ± 2.5 mm/yr; DeMets, 1995), the rate of motion during the Late Miocene phase of northwestern rifting would reduce to ~21 mm/yr. This progressive increase in the rate of separation (7.7–8.3 mm/yr prior to 12 Ma to ~21 mm/yr 12–3.6 Ma) is likely related to the decreasing yield strength of the continental lithosphere as it is mechanically and thermally thinned and, eventually, completely severed by the onset of transtension.

Our reconstruction for the original position of Baja California relative to mainland Mexico ca. 30 Ma (Fig. 15) approximates those of Fletcher et al. (2007) and Sutherland et al. (2012) at 12.5–14 Ma. Those reconstructions implicitly imply no extension prior to 14 Ma and a high rate of separation of 32–35 mm/yr, uncommon for continental rifts where geologically measured rates of opening are typically <10 mm/yr (Calais et al., 1998; Fernandes et al., 2004; Bendick et al., 2006). An implication of our reconstruction of ~475 km of northwest motion of Baja California since ca. 30 Ma is that little dextral shear is required to be accommodated within the Gulf of California after 12.5 Ma other than the ~275–300 km estimated from the correlation of geologic units (Oskin et al., 2001; Miller and Lizarralde, 2013). However, according to global plate circuit reconstructions, since 12.3 Ma, the Pacific plate has moved ~600 km to the northwest relative to stable North America at the latitude of southern Baja California (Atwater and Stock, 1998). This implies that ~300 km of additional oblique shear must have been accommodated since that time between the Pacific plate and stable North America. One possibility is that this deformation occurred in the belt of faults on the western margin of Baja California (Tosco-Abreojos and San Benito faults) that currently accommodates 10% of the Pacific–North America relative motion (Plattner et al., 2009). This possibility would agree with an early estimation of right-lateral motion along this fault belt (Stock and Hodges, 1989), but contrasts with the more limited motion estimates by Fletcher et al. (2007) through correlation of the Magdalena Fan (Fig. 1) with its most probable source. Another place where Late Miocene extension may have been accommodated is in the Basin and Range east of the SMO unextended core. Henry and Aranda-Gomez (2000) reported an extensional episode in this wide area (Fig. 1) between 12 and 6 Ma, although this is not quantified. A third possibility is that deformation has been underestimated in the GEP and some additional motion may be distributed within the Gulf of California in faults buried in the wide coastal plain and shelf of the Nayarit and Sinaloa margin. All these options may contribute to account for the missing ~300 km of displacement between the Pacific and North America plates, but more structural, geochronologic, and geophysical data are needed to assess their respective balance.

**Genesis of Magmatism**

The significant pre–Late Miocene extensional deformation documented in this work for western Mexico has important implications for the genesis of magmatism in the Gulf of California region. The overall calc-alkaline geochemical character, relatively primitive isotopic signature, and the suprasubduction position of the SMO and the Comondú volcanic rocks have traditionally led to interpretations of these provinces as being the manifestation of arc volcanism, whereas volcanism related to the rifting process would have appeared only after subduction ceased. This view needs to be revised in light of the results presented here and elsewhere (see Bryan et al., 2008, 2013). As we have shown, extension and crustal thinning spatially associated with volcanism occurred for at least 15 m.y. prior to the end of subduction.

Geologically observable extension began in the southern SMO just after the major Oligocene ignimbrite flare-up. For example, the ca. 24 Ma El Salto–Espinazo ignimbrite succession...
Early extension in the Gulf of California

filled a tectonic depression bounded by this early normal faulting. Extension continued during the Early Miocene, likely triggering the last major sequence of ignimbrite eruptions ca. 21–20 Ma (Nayar ignimbrite succession). At the same time, volcanism became bimodal and was characterized by more effusive activity. A belt of rhyolitic domes was emplaced along north-northwest–trending normal faults that bound extensional basins localized along the site of the opening of the Gulf of California. Mafic lavas and rhyolitic domes filled major grabens along the coast of Sinaloa. Geochemistry of Early Miocene silicic volcanism as well as the zircon antecrystic signature of many samples is consistent with significant crustal assimilation and melting induced by the arrival of mafic magmas in the crust. Similarly to Sinaloa, in southern Baja California the middle member of the Comondú Group was likely emplaced in actively extending tectonic basins, as suggested by the dominance of volcanioclastic sediments with respect to primary volcanic materials (Dorsey and Burns, 1994; Umhoefer et al., 2001; Drake, 2005). The dacitic-andesitic nature of the volumetrically modest Comondú volcanism has alternatively been explained by mixing of basaltic and rhyolitic magmas rather than by fluid fluxing of the mantle wedge above the subducting Guadalupe and Magdalena plates (Bryan et al., 2013). By ca. 16 Ma, the subducting plate at the trench may have been as young as 3–5 Ma (Ferrari et al., 2012, Fig. 18 therein) and thus too dry and hot to release any meaningful amount of fluids once it reached the appropriate depth to flux the mantle wedge (i.e., 105 km; Syracuse and Abers, 2006; see also Peacock and Wang, 1999). In this scenario, melting of sublithospheric mantle was essentially driven by decompression, induced by the significant lithosphere thinning going on at this time. The apparent subduction-related signature of SMO and Comondú volcanics is most likely a feature inherited by the previous >100 m.y. history of subduction, and was acquired by lithospheric assimilation and melting. It is interesting that our easternmost Early Miocene basaltic sample (TS 16, 24 Ma) has an intraplate signature (albeit with some crustal assimilation), suggesting that by that time, at least beneath the eastern part of the SMO, the mantle was already devoid of any subduction influence. We conclude that since the Late Oligocene, crustal extension and magmatism were intrinsically linked and decompression melting progressively overwhelmed any flux melting of the mantle wedge. A widespread mafic pulse of volcanism occurred ca. 12–10.5 Ma along the Sinaloa and Nayarit coasts, as well as on the conjugate margin of Baja California and in the rifted continental blocks submerged in the Gulf of California. Although this volcanism postdates the end of subduction off Baja California, it still shows a variable subduction signature and a striking difference with the OIB-like late Pliocene to Pleistocene basalts emplaced in the same areas. The regional distribution of this volcanism over a 700-km-long belt from Sinaloa to Nayarit in a narrow time frame points to a common mechanism of mantle melting, albeit modulated by different degrees of fractional crystallization and crustal assimilation. We propose that this mafic volcanism is related to the acceleration of the rate of separation between Baja California and mainland Mexico once subduction ended and the peninsula started to be dragged northwest by the Pacific plate. The increasing northwestward motion of Baja California follows its coupling with the shallow part of the subducted Magdalena plate once the lower part of the slab had detached and started foundering in the mantle (Ferrari, 2004). The precise location of slab detachment in the Gulf of California area is debated, but it is thought to have initiated 13–12 Ma (Calmus et al., 2003; Pállares et al., 2007; Brothers et al., 2012). The ascent of hot asthenospheric mantle into the slab gap that formed after the slab detachment may have triggered extensive melting of the subduction-modified former mantle wedge, which ultimately led to the ca. 12–10.5 Ma mafic pulse. However, the scrubbing of a subduction signature from the mantle and the consequent melting of undeposited asthenosphere in the Nayarit-Sinaloa margin was delayed until the late Pliocene, whereas calc-alkaline melts are still being generated in some locations in eastern Baja California.

CONCLUSIONS

The data presented in this work show that the GEP represents the last and most visible phase of rifting in the Gulf of California driven by the oblique divergence of the Baja California peninsula from the Mexican mainland. Although there are different views on how oblique divergence was partitioned between the San Benito–Tosco–Abreojos fault system and the Gulf of California on the two sides of the peninsula (Stock and Hodges, 1989; Gans, 1997; Oskin and Stock, 2003; Fletcher et al., 2007), there is no doubt that this last episode of rifting and its associated volcanism was primarily controlled by plate boundary forces, namely the progressive dragging of Baja California to the northwest by the Pacific plate and the upwelling of asthenosphere into the slab gap produced by the detachment of the lower part of the subducted slab.

The causes of the early Basin and Range–type extension, however, are not completely understood. The onset of extension in western Mexico was considered to be Oligocene in the eastern part of the SMO and to have subsequently migrated to the west to reach the Gulf of California area at the end of the Middle Miocene (Cameron et al., 1989; McDowell and Maugé, 1994; Henry and Aranda-Gómez, 2000; Luhr et al., 2001; Umhoefer et al., 2001; Aranda-Gómez et al., 2003). By contrast, our work indicates that a sustained period of lithospheric stretching affected the entire western Mexico, from the eastern part of the SMO to the region of the future Gulf of California, since the Late Oligocene. The first part of this extensional history, producing the Mexican Basin and Range, occurred during the last period of subduction beneath North America, and broadly coincided with the silicic volcanism of the SMO and the intermediate volcanism of the Comondú Group. The initial wide zone of rifting between ca. 30 and 20 Ma subsequently focused in the westernmost part of the SMO and the Gulf of California region ca. 20–18 Ma to form a narrower rift where the Comondú Group was being deposited. Subduction of the Magdalena microplate started waning by 15 Ma, and ultimately ceased by ca. 12–11.5 Ma, to be replaced by oblique divergence (Lonsdale, 1991; Tian et al., 2011). Therefore, a significant part of extension in the southern Gulf of California (the proto–Gulf of California) occurred prior to subduction termination and cannot be directly associated with the interaction between the Pacific and North America plates.

Previous studies (e.g., Ferrari et al., 2002, 2007; Bryan et al., 2008, 2013), confirmed by this work, showed that the outbursts of silicic volcanism in the SMO (the ignimbrite flares) cannot be explained by an assimilation and fractional crystallization process alone, but were dominantly produced by intrusion of large amount of mafic melt in the crust. The accumulation of mafic magma in the crust had to be sufficiently rapid to produce enough silicic melt to feed many large-volume ignimbrite eruptions in a short time span (typically 1–2 m.y.). The occurrence in the SMO of two ignimbrite flares suggests that mantle melting and the production of mafic magmas also occurred in two main pulses, ruling out a steady-state flux melting process related to a normal subduction regime.

Silicic magmatism and extension in the U.S. part of the Basin and Range province has been attributed to boundary forces (i.e., trench retreat, rollback, and steepening of the subducting Farallon plate; Dickinson and Snyder, 1978; Best and Christiansen, 1991; Ward, 1991; Bohannon and Parsons, 1995; Dickinson, 2002; McQuarrie and Oskin, 2010) or body forces (i.e., active rifting induced by upwelling of hot asthenosphere.
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within a slab window or gravitational collapse of a thickened crust; e.g., Houseman et al., 1981; Gans et al., 1989; Harry et al., 1993; Lee- man and Harry, 1993; Axen et al., 1993; Wilson et al., 2005; Wong et al., 2010). As shown by Sonder and Jones (1999), a single mechanism is unable to explain all the extensional deformation of the western United States and a combination of both groups of forces is required. The cause of the Basin and Range extension and the ignimbrite fl are-ups in the SMO is probably related to a similar combination of processes, but assessing the role and weight of each factor needs a more complete data set on fault activity, magmatic evolution, and crustal structure.

We conclude that, in the past 30 m.y., western Mexico has been dominated by lithospheric extension that was produced by different geodynamic mechanisms. The real change ca. 12.5 Ma was the direct interaction between the Pacific and North America plates that resulted in rifting focusing on the westernmost part of a wide and already thinned belt of lithosphere and in changing the kinematics of deformation, imposing a high degree of obliquity. In this context, the GEP can be only distinguished from the previous episodes of extension by its right-lateral component of deformation and thus should be more properly called the Gulf transtensional province.

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APPENDIX 1. U-Pb LA-ICP-MS DATING METHODOLOGY

U-Pb ages on separate zircons were obtained by laser ablation–inductively coupled plasma–mass spectrometry (LA-ICP-MS) at the Laboratorio de Estudios Isotópicos, Centro de Geociencias, Universidad Nacional Autónoma de México, according to the procedures reported in Solari et al. (2010). The analytical data are reported in the Supplemental File [see footnote 1].

The Plešovice reference zircon (ca. 337 Ma; Sláma et al., 2008) was used in combination with NIST 610 standard glass to correct for instrumental drift and down-the-matrix fractionation and to recalculate elemental concentrations, using folsite software (Paton et al., 2010) in combination with the VisuAge data reduction scheme of Petrus and Kamber (2012). Precision on the measured 206Pb/238U, 207Pb/235U, and 208Pb/204Pb ratios was typically <0.95%, 0.7%, and 1.1% for relative standard deviation, respectively. Replicate analyses of the Plešovice zircon (Sláma et al., 2008) indicate an external reproducibility of 0.95%, 0.7%, and 1.6% on the measured 206Pb/238U, 207Pb/235U, and 208Pb/204Pb ratios, respectively. These errors are quadratically included in the quoted uncertainties for individual analyses of the analyzed zircons. Because its signal is swamped by the 204Hg contained in the carrier gases, 206Pb was not analyzed during this study. When needed, the common Pb correction was performed, employing the algebraic method of Andersen (2002). With this method, the analysis of Cenozoic zircons is a complex task, due to the tiny signal of 206Pb, which yields something imprecise (i.e., discordant) analyses, difficult to correct with the Andersen (2002) algorithm. The concordance filter is therefore not applied, because this would eliminate most of the obtained ages. The concordia, weighted mean ages, as well as age error calculations, were performed using Isoplot v. 3.70 (Ludwig, 2008). The 206Pb/238U ages are preferred for grains younger than 1000 Ma because of the uncertainty involved in determining the 206Pb isotope in young crystals.

APPENDIX 2. 4Ar/3Ar AR METHODOLOGY

The 29 new ages presented here were obtained over a period of 6 years. All the samples were step-heated. The temperature-controlled Ta furnace was used to analyze the argon of four samples; for these, the argon isotopes were analyzed with a MS10 mass spectrometer. The rest of the samples were step-heated using an argon-ion laser beam and the extracted argon was analyzed with a VG5400 mass spectrometer. With the exception of samples ESC 7, MDC 10, LM 2, PER 12, and PER 14, which were prepared at the mineral separation laboratory of the Centro de Geocienencias, Universidad Nacional Autónoma de México (CGEO-UNAM), the samples were prepared at Departamento de Geología, Centro de Investigación Científica y de Educación Superior de Ensenada (CICESE). The basic sample preparation consisted of crushing and overnight at ~60° C. The rock fragments or mineral groundmass fragments were free of phenocrysts. The groundmass or mineral crystals selected for analysis were generally ~500 µm; however, some samples were selected to be separated or where it could be observed that the fragment size was closest to the size of the minerals to be separated or where it could be observed that the groundmass fragments were free of phenocrysts. Mineral-separation procedure consisted of magnetic Frantz separation, heavy liquids if necessary, and final examination of the samples under the microscope to ensure >99% purity of the concentrate. For some samples, no mineral concentrate could be separated; these samples were examined under the microscope and the phenocrysts were eliminated by hand; these were labeled groundmass. If no phenocrysts were observed, the sample was labeled whole rock.

The reference minerals and minerals were irradiated in position 5C and sample ES 10 was irradiated in position 8C. The irradiations were performed in the U-enriched research reactor of McMaster University in Hamilton, Ontario, Canada. Because the project was conducted over several years, the samples were irradiated in different groups. Samples DANA 46a and ES 10 received 40 MWh; the rest of the samples received 30 MWh. With the exception of sample DANA 46a, which was irradiated without a Cd liner, rock fragments were covered with a Cd liner to block thermal neutrons. The irradiation monitors used were: TCR-2 (split G93) sanidine (28.34 ± 0.28 Ma; Renne et al., 1998); PCT-2 sanidine (29.201 ± 0.046 Ma; Kuiper et al., 2008); HD-B1 biotite (24.18 ± 0.09 Ma; Schwarz and Triefelt, 2007); CATAV 7–4 biotite (89.13 ± 0.35 Ma; internal standard calibrated with hornblende lb 3gr at the University of Toronto, with hornblende MMhb 1 at the University of Nice and at CICESE with sandine TCR-2, sanidine PCT-2, biotite HD-B1, and 128.1 Ma biotite LP-6; Roddick, 1983). The irradiation monitors were distributed among the samples. The irradiation monitors were fused in one step to calculate J. The J value used for the samples came from the monitor that was closest to the sample during irradiation. And the argon experiments were preceded by a blank measurement of all the argon masses. Upon blank subtraction, the argon isotopes were corrected for mass discrimination, calcium, potassium, and chlorine neutron-induced interference reactions. The parameters used to correct for neutron-induced interference reactions were: (40Ar/36Ar)Ca = 6.51 × 10–2; (40Ar/39Ar)Ca = 2.54 × 10–2; (40Ar/37Ar)Ca = 8.7 × 10–5 for DANA 46a. For the rest of the samples, which had Cd liners, the parameters were: (40Ar/36Ar)Ca = 6.50 × 10–4; (40Ar/39Ar)Ca = 2.55 × 10–4; (40Ar/37Ar)Ca = 0. Mass 36 was also corrected for chlorine-derived “Ar (n, γ) 37Cl → Ar + β with a half life of 3.1 × 1010 yr. Isotopes 39Ar and 39Ar were corrected for radioactive decay. The constants recommended by Steiger and Jäger (1977) were used in all the calculation, while all the straight line calculations were performed with the equations presented in York et al. (2004). All errors are reported at 1σ level. The errors in the integrated, plateau, isochron, and weighted mean age include the uncertainty in the J parameter. In addition, for the plateau, isochron, and weighted mean ages, the goodness of fit was included in the age uncertainty whenever the mean square of weighted deviates was >1. The integrated ages were calculated adding the fractions of the step-heating experiments. Plateau ages were calculated with the weighted mean of three or more consecutive fractions, which were in agreement within 1 σ errors excluding the uncertainty in J. All the data were plotted on an Ar/Ar correlation diagram to determine the composition of the (Ar/Ar) of the samples. Tables with the relevant Ar/Ar data of all the experiments are given here. A brief discussion of the Ar/Ar results follows, tables with the relevant Ar/Ar data of all the experiments and the figures with age spectrum, Ar/Ar, Ar/Ar diagram, and the Ar/Ar correlation diagram for each sample are given in the Supplemental File [see footnote 1].

APPENDIX 3. SAMPLE PREPARATION FOR TRACE ELEMENT ANALYSIS

Mafic to intermediate samples were analyzed using the procedure described in Mori et al. (2007). For silicic samples (SiO2 > 60%), two additional digestion steps were carried out, in order to achieve complete dissolution of refractory minerals (e.g., zircon). The same amount of sample (50 mg) was weighed in 15 mL Teflon vials, and after an initial overnight attack
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with 1 mL HF and 0.5 mL 8 N HNO₃ on a hot plate at 90 °C, followed by evaporation to almost dryness, samples were transferred to 1.5 mL Teflon vials with 1 mL HF and 0.5 mL 8 N HNO₃. Closed vials were then placed inside the Teflon liner of metal-jacketed Parr pressure bombs, to which 3 mL of a 2:1 HF-8N HNO₃ mixture were added in order to equalize the pressure inside and outside the vials and prevent solution loss. The bomb was sealed and heated in an oven at 190 °C for 5 days. Thereafter, the bomb was allowed to cool to room temperature and the vials were removed, opened, and placed on a hot plate to evaporate the acids to almost dryness. After addition of 1.5 mL HCl, the vials were placed again in Parr bombs and 3 mL of 6N HCl were added to the Teflon liner. The bombs were heated at 190 °C in an oven for 24 h. This step is required in order to decompose the fluoride that might have formed in the previous steps. After cooling, vials were removed from the bombs and the samples were transferred to the original 15 mL vials, which have a larger surface area that allows faster evaporation. The final steps consisted in evaporating the samples to dryness, fluxing the samples twice with 16N HCl, and letting the samples to dryness after each step. Then, 2 mL deionized water and 2 mL 8N HNO₃ were added to the samples and the closed vials were left overnight on a hot plate at 90 °C. For analysis, the samples were diluted by weight to 100 g (1:2000 dilution) with an internal standard solution containing Ge (10 ppb), In (5 ppb), Tm (5 ppb), and Bi (5 ppb). Calibration was performed with the international rock standards AGV-2, BHVO-2, BCR-2, J-2, and JR-1. Sample DANA 46a, ROCA 3J 4a, and ROCA 24J 33 were analyzed by X-ray fluorescence spectrometry and by inductively coupled plasma–mass spectrometry at the GeoAnalytical Laboratory of Washington State University with the methods described in Castillo et al. (2010).
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