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Paleomagnetic constraints on the kinematic relationship between the Guerrero terrane (Mexico) and North America since Early Cretaceous time

Lydian M. Boschman[1,†], Roberto S. Molina Garza[2,†], Cor G. Langereis[1,†], and Douwe J.J. van Hinsbergen[1,†]
[1]*Department of Earth Sciences, Utrecht University, Heidelberglaan 2, 3584 CS Utrecht, Netherlands*
[2]*Centro de Geociencias, Universidad Nacional Autónoma de México, Boulevard Juriquilla 3001, Juriquilla, Querétaro 76230, Mexico*

ABSTRACT

The North American Cordillera has been shaped by a long history of accretion of arcs and other buoyant crustal fragments to the western margin of the North American plate since early Mesozoic time. The southernmost accreted terrane is the Guerrero terrane of southwestern Mexico, a latest Jurassic–Cretaceous volcanic arc built on a Triassic accretionary prism. Interpretations of the origin of the Guerrero terrane vary: Some authors consider it a far-traveled, exotic intra-oceanic island arc, while others view it as the (par)autochthonous, extended North American continental margin. We present new paleomagnetic and U-Pb zircon data from Lower Cretaceous sedimentary rocks of the Guerrero terrane. These data show that the Guerrero terrane has a latitudinal plate-motion history equal to that of the North America plate, both before and after accretion. This confirms paleogeographic models in which the Guerrero arc successions formed on North American crust that rifted away from the Mexican mainland by approximately east-west opening of a back-arc basin above an eastward-dipping subduction zone. Additionally, it renders alternative paleogeographical models in which the Guerrero terrane is considered to be exotic to the North American continent unlikely. The phase of back-arc spreading resulted in the short-lived existence of an additional "Guerrero" tectonic plate between the North American and Farallon plates and, upon closure of the back-arc basin, the growth of the North American continent.

†l.m.boschman@uu.nl, rmolina@geociencias.unam.mx, c.g.langereis@uu.nl, d.j.j.vanhinsbergen@uu.nl.

INTRODUCTION

The North American Cordillera has been shaped by a long history of accretion of arcs and other buoyant crustal fragments to the western margin of the North American plate. Adding to the complexity of interpreting the plate-tectonic provenance of these accreted terranes, the continental margin has also undergone episodes of extension associated with the opening of forearc and back-arc basins, periods of shortening producing inversion of preexisting basins and fold-and-thrust belts, and translational deformation (moving terranes north along major strike-slip faults; Nokleberg, 2000; Johnston, 2001; McQuarrie and Wernicke, 2005; Umhoefer and Blakey, 2006; Wyld et al., 2006; DeCelles et al., 2009; Schellart et al., 2010; Hildebrand, 2013, 2015; Shephard et al., 2013; Sigloch and Mihalynuk, 2013). Over the last decades, contrasting paleogeographic models have been proposed to explain the assembly and evolution of the North American Cordillera. In a first group of models, the North American plate collided with either an offshore "Cordilleran ribbon continent" (e.g., Johnston, 2001; Hildebrand, 2013) or a sequence of exotic archipelagos and subduction complexes (e.g., Sigloch and Mihalynuk, 2013) via westward subduction. A second group of models considers the Cordillera to have formed as the result of long-lived eastward subduction of Panthalassa/Pacific lithosphere below the North American continent (e.g., Monger et al., 1982; Nokleberg, 2000; Monger and Price, 2002; Scotese, 2004). In these models, the transfer of exotic crustal material from the subducting oceanic plates to the overriding plate played a minor role, and deformation within the overriding North American plate accounted for most of the formation of the Cordilleran.

From a plate-tectonic perspective, terrane accretion and deformation of the North American Pacific margin have been governed by the interaction between the continental North American plate in the east and the oceanic plates of the Panthalassa (and later, Pacific) Ocean in the west (Engebretson et al., 1985). Due to subduction and the consequent loss of oceanic spreading records, reconstructions of the plate-tectonic configuration of the Mesozoic northeast Pacific are prone to large uncertainties. However, marine magnetic anomalies from the northern part of the Pacific plate are interpreted to have formed conjugate to now-subducted oceanic lithosphere of the conceptual Izanagi, Farallon, and Kula plates, which were located north, east, and northeast of the Pacific plate, respectively. Relative to North America, the Farallon and Kula plates moved toward the northeast, with a larger northward motion component for the Kula plate than for the Farallon plate (Doubrovine and Tarduno, 2008; Wessel and Kroenke, 2008; Seton et al., 2012; Wright et al., 2016). Magnetic seafloor data cannot resolve whether the Farallon and Kula plates completely filled the northeastern Panthalassa basin, extending to the western margin of North America, or if other, smaller plates existed between the major oceanic Panthalassa plates and the continent. Reconstruction of the complex plate-tectonic history of the northeastern Panthalassa basin and the Cordillera further relies on quantitative plate-motion estimates from geological observations and paleomagnetic data, providing the kinematic underpinnings for paleogeographical models.

This study focuses on the southernmost of the accreted Cordilleran terranes: the Guerrero terrane of southwestern Mexico. This terrane consists of an Upper Triassic metasedimentary basement, unconformably overlain by Middle Jurassic–Upper Cretaceous arc assemblages (Centeno-García et al., 2003, 2008; Talavera-Mendoza et al., 2007; Martini et al., 2011). The eastern boundary of the Guerrero terrane

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is represented by a deformed belt that includes Tithonian–Aptian, highly sheared and folded meta-turbidites interbedded with uppermost Jurassic–lowermost Cretaceous rhyodacitic dikes and lava flows and Aptian intraplate-like and mid-ocean-ridge basalts (Tardy et al., 1994; Freydier et al., 1996; Mendoza and Suastegui, 2000; Martini et al., 2011, 2014). The presence of mid-ocean-ridge basalts indicates that during Aptian time, the Guerrero arc was located on a tectonic plate other than, and separated by a spreading ridge from, the North American plate.

Both categories of paleogeographic models described in the preceding paragraphs have been used to interpret the origin of the Guerrero terrane. On the one hand, the Guerrero terrane is interpreted as an exotic intra-oceanic island arc that accreted to mainland Mexico by closure of a large oceanic basin via westward subduction (Lapierre et al., 1992; Tardy et al., 1994; Freydier et al., 1996; Dickinson and Lawton, 2001; Umhoefer, 2003; Hildebrand, 2013; Sigloch and Mihalynuk, 2013). This interpretation predicts the presence of an additional oceanic plate and at least one subduction zone between the Guerrero terrane and the North American plate. Depending on the paleogeographic model, the Guerrero terrane was either located on the eastern margin of the Farallon plate (Dickinson and Lawton, 2001; Umhoefer, 2003; Sigloch and Mihalynuk, 2013), predicting latitudinal (northward) motion of the Guerrero terrane relative to the North American plate prior to its accretion, or in a similar position on an undefined plate within the Panthalassa Ocean (Lapierre et al., 1992; Tardy et al., 1994; Freydier et al., 1996; Hildebrand, 2013).

On the other hand, the Guerrero arc successions are interpreted to have formed on North American crust that rifted away from the Mexican mainland by opening of a back-arc basin above an eastward-dipping subduction zone (Cabral-Cano et al., 2000; Elías-Herrera et al., 2000; Centeno-García et al., 2008, 2011; Martini et al., 2011; Fitz-Díaz et al., 2017). Plate tectonically, this interpretation includes ongoing eastward subduction below North America and the temporal existence of a divergent plate boundary between a small Guerrero plate and the North American plate. Predicted plate motion of this Guerrero plate relative to the North American plate would have been limited, and would have occurred more or less perpendicular to the approximately N-S–oriented trench and ridge; i.e., such plate motions would have been mainly longitudinal.

To determine the plate-tectonic provenance of the Guerrero terrane, quantitative kinematic data are key. To this end, we constrained the latitudinal plate motion of the Guerrero terrane using paleomagnetic data. High-quality paleomagnetic data from the Guerrero terrane are available from Upper Cretaceous to recent stratigraphic units but are not available for older intervals. Here, we present new paleomagnetic data and a U-Pb detrital zircon age from Lower Cretaceous sedimentary rocks of the Guerrero terrane.

GEOLOGICAL SETTING

The pre-Cenozoic basement of Mexico is subdivided into multiple tectonic terranes constituting two groups: the Proterozoic eastern terranes, which accreted to the southern part of the North American craton in late Paleozoic time (Keppie, 2004; hereafter: "mainland Mexico"), and the western accreted terranes, which together form the Guerrero (super- or composite) terrane (Dickinson and Lawton, 2001; Centeno-Garcia et al., 2011; see Fig. 1). The Guerrero terrane consists mostly of submarine and locally subaerial volcanic and sedimentary rocks intruded by plutons that range in age from Tithonian (Late Jurassic) to Cenomanian (Late Cretaceous; Talavera-Mendoza et al., 2007; Centeno-García et al., 2008; Martini et al., 2011). Underlying these arc successions, there is the Arteaga complex, consisting of Triassic greenschist-facies meta-turbidites, which have been interpreted as an accretionary prism (Ranson et al., 1982; Cuevas-Pérez, 1983; Monod and Calvet, 1991; Centeno-García et al., 1993; Centeno-García and Silva-Romo, 1997; Talavera-Mendoza et al., 2007). This basement complex contains detrital zircons that have returned Grenville, Pan-African, and Permian U-Pb ages that match the signature of mainland Mexico and the northwestern margin of South America (Centeno-Garcia et al., 2011). Intruding and overlying the Arteaga complex, there are Middle to Upper Jurassic arc granitoids, submarine rhyolitic lavas, and volcaniclastic rocks (Bissig et al., 2003; Centeno-García et al., 2003). These Jurassic arc rocks were deformed and exhumed prior to onset of the major phase of arc volcanism in the Tithonian (Centeno-García et al., 2003, 2008). The eastern boundary of Guerrero exposes sheared and folded meta-turbidites interbedded with volcanic rocks, interpreted as remnants of an oceanic basin (the Arperos Basin) that separated the Guerrero terrane from the Mexican mainland (Tardy et al., 1994; Freydier et al., 1996; Mendoza and Suastegui, 2000; Martini et al., 2011, 2014). This deformed belt exposes Tithonian–Barremian felsic dikes and lava flows containing Paleozoic and Precambrian inherited zircons, as well as Aptian–Cenomanian intraplate-like and mid-ocean-ridge basalts (Elías-Herrera et al., 2000; Mortensen et al., 2008; Martini et al., 2011, 2014). This suggests that during the Tithonian–Barremian, the Arperos Basin was floored by sediments derived from continental crust, or by continental crust itself, and oceanic crust formed since the Aptian. Accretion of the Guerrero terrane to mainland Mexico may have been diachronous, as a basaltic lava flow from the southeastern part of the Guerrero terrane suture belt, interpreted to reflect ongoing back-arc spreading, has been dated at 93 Ma (Ar-Ar age; Elías-Herrera et al., 2000), whereas an undeformed sedimentary overlap assemblage on the central part of the Guerrero terrane suture belt has yielded Albian paleontologic ages (Chiodi et al., 1988). After accretion, arc magmatism continued throughout western Mexico in the Sierra Madre del Sur and, since the middle Miocene, in the Trans-Mexican volcanic belt (Morán-Zenteno et al., 2007; Ferrari et al., 2012).

SAMPLING LOCATIONS

To resolve the absence of high-quality paleomagnetic data from the early history of the Guerrero terrane, we extensively sampled two Lower Cretaceous sedimentary sections. In Michoacán State, near de town of Zitácuaro (locality AB, Fig. 1), we collected 133 samples from a sedimentary succession consisting of gray fine-grained sand, shale, limestone, and some monomictic conglomerate containing limestone clasts. The samples were collected from fine-grained limestone intervals only. These limestone beds contain Upper Aptian ammonites such as *Dufrenoya* sp., *Huastecoceras* sp., and *Caseyella* sp., as well as the planktonic foraminifera *Globigerinelloides ferreolensis* Moullade, *Globigerinelloides aptiensis* Longoria, *Hedbergella trochoidea* (Gandolfi), and *Hedbergella roblessae* (Obregón de la Parra; C. González, 2013, personal commun.). This association corresponds to the Upper Aptian (ca. 119–112 Ma) *Globigerinelloides ferrolensis* fauna of Premoli Silva and Verga (2004).

In Jalisco State, near the town of Tamazula de Gordiano (locality AL, Fig. 1), we collected 168 samples of reddish andesitic sandstones and shales at five sites (AL1–AL5) along two parallel roads ~8 km apart. These redbeds were mapped by Romo Ramírez et al. (2001) as the Valanginian–Hauterivian Alberca Formation and interpreted as the stratigraphically lowest volcaniclastic unit of the Guerrero arc succession in this region. However, the age of the Alberca Formation is not well constrained, as Berriasian–Hauterivian and Barremian–Cenomanian ages (Cuevas [1981] and Grajales and López [1984] *as cited in* Centeno-García et al., 2008) have also been reported. At site AL4, stratigraphically high in the sampled succession,

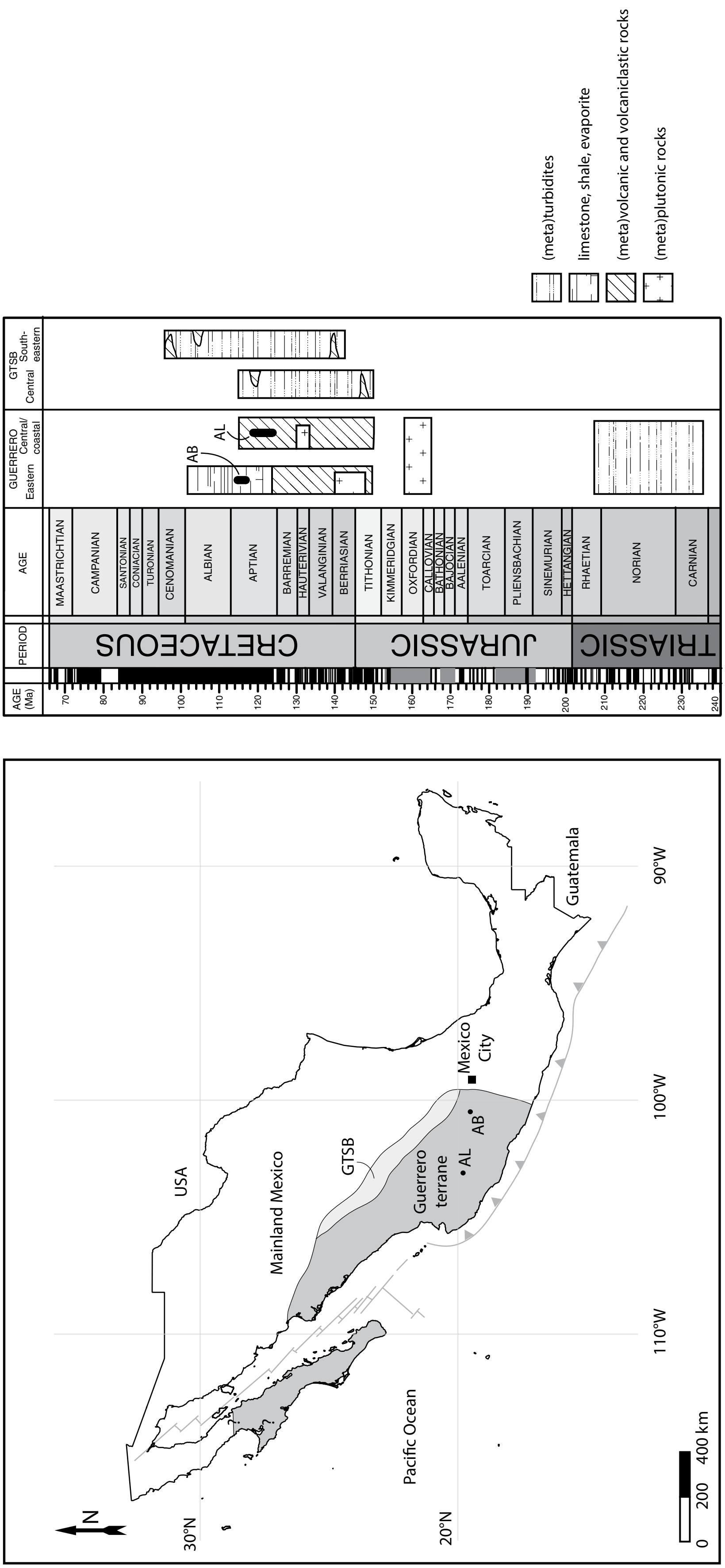


Figure 1. (A) Simplified geological map of the Guerrero terrane and the Guerrero terrane suture belt (GTSB), modified from Ortega-Flores et al. (2014), including sampling locations AL and AB. (B) Simplified stratigraphic column of the Guerrero terrane and the Guerrero terrane suture belt, modified from Centeno-García et al. (2008) and Martini et al. (2014).

we took a sample for U-Pb detrital zircon dating to obtain a maximum depositional age.

METHODS

Paleomagnetic cores with a diameter of 2.5 cm were sampled with a gasoline-powered motor drill, and their orientation was measured with a magnetic compass with an inclinometer attached. The cores were cut into samples of 2.2 cm length using a double-blade circular saw. Laboratory analyses were carried out at the Paleomagnetic Laboratory Fort Hoofddijk of Utrecht University, Netherlands. Thermomagnetic analyses to determine the nature of magnetic carriers were performed on representative samples for each site, using a horizontal translation-type Curie balance with a cycling applied magnetic field, usually 100–300 mT (Mullender et al., 1993). Several heating-cooling cycles were applied to detect magneto-mineralogical alterations during heating. We used the following temperature scheme (in °C): 20–150, 50–250, 150–350, 250–400, 300–450, 350–500, 400–700. Furthermore, all 301 samples were subjected to either stepwise thermal (TH) or alternating field (AF) demagnetization, and the natural remanent magnetizations were measured on a 2G DC SQUID cryogenic magnetometer. Demagnetization steps used were 4, 8, 12, 16, 20, 25, 30, 35, 40, 45, 50, 60, 70, 80, 90, and 100 mT for AF treatments and variable temperature increments of 20–50 °C up to 680 °C for TH treatments. Demagnetization diagrams were plotted on orthogonal vector diagrams (Zijderveld, 1967), and the magnetic components were determined via principal component analysis (Kirschvink, 1980). The fold test (Tauxe and Watson, 1994) and the bootstrapped coordinate reversal test (Tauxe, 2010) were used when applicable. We calculated site mean directions using Fisher (1953) statistics on virtual geomagnetic poles following statistical procedures described in Deenen et al. (2011). A 45° cutoff was applied to the virtual geomagnetic poles per locality (Johnson et al., 2008). Sample interpretation and statistical analysis were conducted using the online portal Paleomagnetism.org (Koymans et al., 2016). All results can be imported into the portal from data files in the Data Repository.[1]

[1]GSA Data Repository item 2018028, 1. Table DR1: Zircon ages per grain, 2. Table DR2: Paleomagnetic data of compiled database, 3. site_locations_DR.kmz: GPS coordinates paleomagnetic sampling sites, 4. pmag_data_DR.pmag: paleomagnetic data for implementation in paleomagnetism.org, is available at http://www.geosociety.org/datarepository/2018 or by request to editing@geosociety.org.

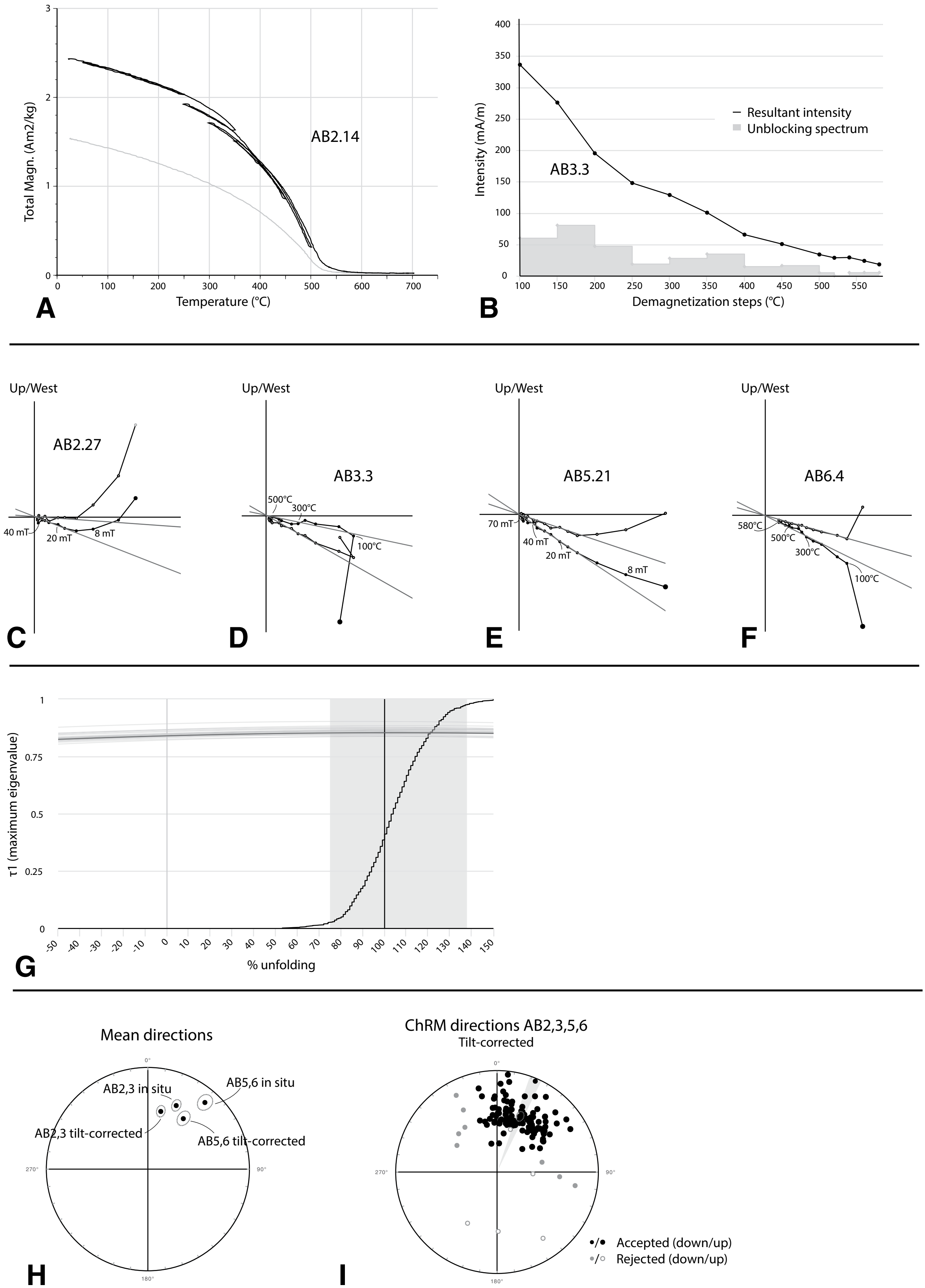

Total Magn. (Am2/kg)
Temperature (°C)
AB2.14
A
Intensity (mA/m)
Demagnetization steps (°C)
Resultant intensity
Unblocking spectrum
AB3.3
B
Up/West
AB2.27
40 mT
20 mT
8 mT
C
Up/West
AB3.3
500°C
300°C
100°C
D
Up/West
AB5.21
70 mT
40 mT
20 mT
8 mT
E
Up/West
AB6.4
580°C
500°C
300°C
100°C
F
τ1 (maximum eigenvalue)
% unfolding
G
Mean directions
AB2,3 in situ
AB5,6 in situ
AB2,3 tilt-corrected
AB5,6 tilt-corrected
H
ChRM directions AB2,3,5,6
Tilt-corrected
Accepted (down/up)
Rejected (down/up)
I

TABLE 1. PALEOMAGNETIC RESULTS

Site(s)	*N*	*N45(is)*	*N45(tc)*	In situ *D*	ΔD_x	*I*	ΔI_x	Tilt corrected *D*	ΔD_x	*I*	ΔI_x	*k*	a_{95}	*K*	$A_{95min} < A_{95} < A_{95max}$
AB23	79	76	75	24.1	3.6	32.1	5.5	12.6	4.1	41.5	5.1	21.3	3.6	19.9	2.1 < 3.8 < 5.4
AB56	51	45	45	40.6	4.5	14.8	8.6	35.1	5.4	39.2	6.9	18.7	5.1	19.4	2.6 < 5.0 < 7.5
AB2356	**130**		**117**					**22.5**	**3.5**	**40.5**	**4.4**	**17.8**	**3.2**	**17.4**	**1.8 < 3.2 < 4.1**
AL1,2 (reversed)	38	35	36	124.6	3.1	–9	6	132.8	4.5	–36.5	6.1	24.3	4.9	33.7	2.9 < 4.2 < 8.6
AL2,3,5 (normal)	99	88	88	327.7	3.3	30.5	5	339.6	4	49.8	3.9	23.6	3.2	20.4	2.0 < 3.4 < 4.9
AL1235	**137**		**123**					**330.2**	**3.7**	**46.5**	**3.9**	**17.5**	**3.1**	**16.2**	**1.8 < 3.3 < 4.0**
AL4	27	26		7.7	5.3	40.1	6.7					32	5.1	35	3.3 < 4.9 < 10.5

Notes: *N*—number of demagnetized specimens; N45(is)/N45(tc)—number of specimens that fall within the 45° cutoff in in situ coordinates (is) and after tilt correction (tc); *D*—declination; *I*—inclination. Bold: locality averages.

We collected sandstone sample AL4DZ from site AL4 for laser ablation–inductively coupled plasma–mass spectrometry U-Pb detrital zircon dating. The U-Pb dating analysis was performed at the Laboratorio de Estudios Isotópicos of the Centro de Geosciencias of Universidad Nacional Autónoma de México, following the procedures described by Solari et al. (2010) and Solari and Tanner (2011).

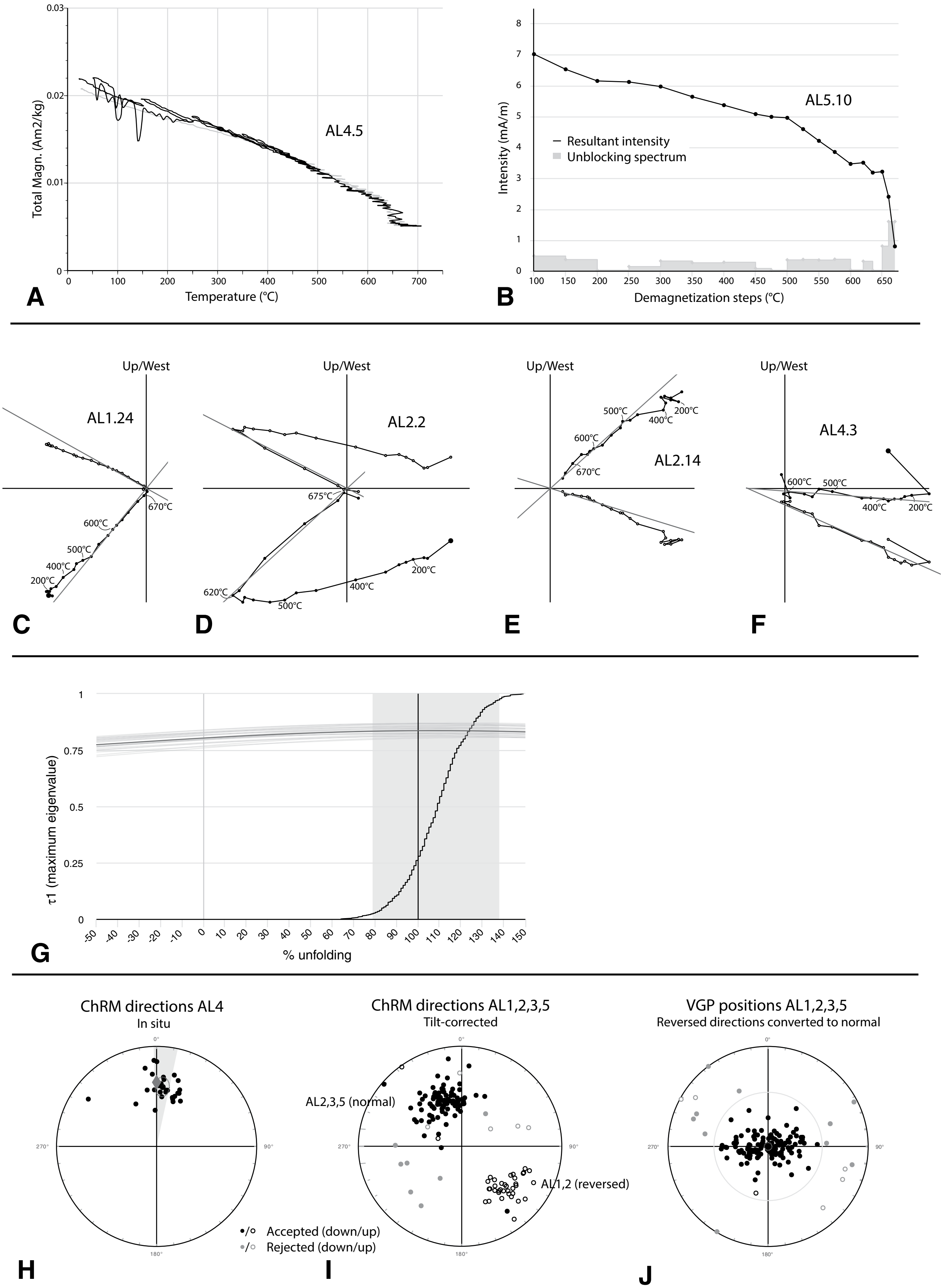
RESULTS

Paleomagnetism

The thermomagnetic curves of the AB limestones are largely reversible, indicating that no alternation occurred during heating; Curie temperatures range 500–560 °C, indicating that the main magnetic carrier is titanomagnetite (Fig. 2A). The intensity decay curve is in line with titanomagnetite (Fig. 2B), which is typical for a volcanic environment. We subjected samples from limestone locality AB to both TH and AF demagnetization treatments and isolated characteristic remanent magnetization (ChRM) values at midrange temperatures (~300–560 °C) or ~16–40 mT (Figs. 2C–2F). Initial intensities ranged 50–800 mA/m. All AB samples yielded normal polarity directions. A slight difference in bedding orientation between AB2-AB3 and AB5-AB6 allowed for a fold test (Fig. 2G), which was positive (optimal clustering between 75% and 138% unfolding). However, AB2-AB3 and AB5-AB6 do not share a common true mean direction (CTMD), because declinations differ significantly between these two groups (Fig. 2H). Inclinations, however, are identical. Due to the resulting artificial elongation in the virtual geomagnetic pole (VGP) position cluster, an inclination shallowing correction could not be performed, but since limestones generally compact insignificantly, this is not problematic. The average ChRM direction (Fig. 2I) is declination (Dec) ± ΔD_x = 22.5° ± 3.5°, inclination (Inc) ± ΔI_x = 40.5° ± 4.4°, n = 117, K = 17.4, A_{95} = 3.2° (Fig. 2G; Table 1). There are four lines of evidence that support the primary nature of the magnetization: (1) Both the in situ and bedding tilt-corrected directions differ significantly from the geocentric axial dipole (GAD) field direction; (2) only normal directions are present, which is consistent with the late Aptian age of the section and magnetization acquisition during the Cretaceous normal superchron; (3) the fold test was positive, supporting pretilt magnetization acquisition; and (4) the distribution of the ChRM values satisfies the quality criteria of representing paleosecular variation (PSV; i.e., A_{95min} = 1.8° < A_{95} = 3.2° < A_{95max} = 4.1°; Deenen et al., 2011). The average ChRM direction results in a paleomagnetic pole at lat. = 68.75°N, long. = 333.4°E, A_{95} = 3.2°, K = 17.4.

Figure 2. Rock magnetic and paleomagnetic results from locality AB. (A) Thermomagnetic curve measured on a Curie balance. Heating in black; cooling in gray. (B) Intensity decay curve. (C–F) Orthogonal vector diagrams for in situ coordinates, where closed (open) symbols indicate declination (inclination). (G) Bootstrapped fold test, with cumulative distribution function (with confidence interval in light gray) based on 1000 bootstraps (average of bootstraps in darker gray). (H) Mean directions for grouped sites AB2-AB3 and AB5-AB6 in both in situ and tilt-corrected coordinates. (I) Characteristic remanent magnetization (ChRM) values for sites AB2, AB3, AB5, and AB6 in tilt-corrected coordinates.

Curie temperatures of the AL red beds are 650–670 °C (Fig. 3A), indicating that the magnetic carrier in the samples is hematite. The maximum applicable alternating field (100 mT) was not high enough to fully demagnetize these hematite-bearing red beds, and so all AL samples were measured by thermal demagnetization up to 680 °C. Initial intensities ranged 8–80 mA/m, and intensity decay curves of the demagnetizations indicate that the red beds contain a mixture of pigmentary and detrital hematite (Fig. 3B). We interpreted a high-temperature component (~500–670 °C) as the ChRM for sites AL1, AL2, AL3, and AL5 (Figs. 3C–3E). Site AL4 only showed a very chaotic component, or no high-temperature component (Fig. 3F), and interpreting a lower-temperature component (~300–550 °C) yielded a direction for in situ coordinates of Dec ± ΔD_x = 5.2° ± 6.7°, Inc ± ΔI_x = 40.0° ± 7.1°, which is statistically indistinguishable from the local GAD field direction (Dec = 0°, Inc = 36°; Fig. 3H). Therefore, we excluded this site from further analysis. Interpreted high-temperature directions from all samples from site AL1 yielded reversed directions, whereas AL3 and AL5 yielded normal directions. AL2 contained both normal and reversed directions, indicating the presence of a reversal within the analyzed sedimentary succession. A fold test (Fig. 3G) on the individual directions of sites AL1, AL2, AL3, and AL5 was positive (optimal clustering with 79%–138% unfolding). The clusters of reversed directions (AL1, part of AL2) and normal directions (part of AL2, and AL3 and AL5; Fig. 3I) did not pass the reversal test. A substantial part of the normal directions is closer to the GAD field direction, so the lack of a positive reversal test is probably best explained by an unresolved normal overprint that influenced the ChRM directions. To diminish this effect, we combined normal and reversed directions (see, e.g., Scheepers and Langereis, 1993), yielding an average direction of Dec ± ΔD_x = 330.2° ± 3.7°, Inc ± ΔI_x = 46.5° ± 3.9°, n = 123, K = 16.2, A_{95} = 3.3° (Fig. 3J; Table 1). Again, there are four lines of evidence that support the primary nature of the magnetization: (1) Both the in situ and bedding tilt-corrected directions differ significantly from the GAD field direction, showing that the ChRM cannot be explained by a recent field; (2) both normal and reversed directions are present, suggesting magnetization acquisition during at least two magnetic polarity chrons; (3) the fold test was positive, supporting pretilt magnetization acquisition; and (4) the distribution of the ChRMs satisfies the quality criteria of representing PSV



Total Magn. (Am2/kg)
Temperature (°C)
AL4.5
A
Intensity (mA/m)
Demagnetization steps (°C)
AL5.10
Resultant intensity
Unblocking spectrum
B
Up/West
AL1.24
670°C
600°C
500°C
400°C
200°C
C
Up/West
AL2.2
675°C
620°C
500°C
400°C
200°C
D
Up/West
AL2.14
500°C
400°C
200°C
600°C
670°C
E
Up/West
AL4.3
600°C
500°C
400°C
200°C
F
τ1 (maximum eigenvalue)
% unfolding
G
ChRM directions AL4
In situ
H
ChRM directions AL1,2,3,5
Tilt-corrected
AL2,3,5 (normal)
AL1,2 (reversed)
Accepted (down/up)
Rejected (down/up)
I
VGP positions AL1,2,3,5
Reversed directions converted to normal
J

(i.e., $A_{95min} = 1.8° < A_{95} = 3.3° < A_{95max} = 4.0°$; Deenen et al., 2011). The cluster of VGP positions is slightly elongated (Fig. 3I), which may indicate inclination shallowing due to compaction (Tauxe and Kent, 2004). However, the elongated shape is in this case largely the result of combining AL1-AL2 (reversed) and AL2, AL3, and AL5 (normal) data with the fanning of the declinations as explained above. Furthermore, detrital hematite is expected to yield a wider range in declinations than can be expected from secular variation, as a result of its mineral properties and mechanism of sedimentation (Tauxe and Kent, 1984; Tan et al., 2002; Iosifidi et al., 2010). For these two reasons, application of the Tauxe and Kent (2004) inclination shallowing correction cannot be justified, as it would significantly overestimate the inclination. We note that compaction cannot be fully excluded, but we argue that since the average inclination of the AB limestones (40.5° ± 4.4°) is replicated by the ~3 m.y. older AL red beds, a significant effect of inclination shallowing is highly unlikely. An inclination shallowing correction with a standard compaction factor of 0.6 (following Torsvik et al., 2012) would increase the average inclination of the red beds to 60.3°. This would predict a southward shift of the Guerrero terrane of ~2000 km within ~3 m.y., which would require unrealistically high tectonic speeds. We therefore used the uncorrected average direction to calculate the paleomagnetic pole at lat. = 62.10°N, long. = 191.81°E, $A_{95} = 3.3°$, $K = 16.2$.

Depositional Age of the AL redbeds

Zircons from sample AL4DZ are subrounded to euhedral in shape and range in length from 80 to 200 μm. Some grains contained inherited cores, but in all cases, we dated the overgrown rims of the grains (Fig. 4C). Of the 111 measured zircon grains, 96 yielded concordant ages (10% cutoff; Table DR1 [see footnote 1]). Even though all ages fell within one group, the large dispersion (mean square of weighted deviation [MSWD] = 6.6) indicates that this group is not one population (Fig. 4B; Spencer et al., 2016). This is in line with the sedimentary nature of the rock, because single populations (associated with in situ zircons) are expected for igneous and metamorphic rocks only (Spencer et al., 2016). To determine the maximum age of deposition, we used the youngest zircon age (youngest single-grain age, following Dickinson and Gehrels, 2009) of 118.8 ± 1.55 (1σ) Ma (Fig. 4B). Sample AL4DZ came from a stratigraphically high level within the sampled succession, and magnetic directions from the upper part of the succession yielded normal directions only, which is consistent with the zircon age overlapping the Cretaceous normal superchron (ca. 125.9–83 Ma; Gradstein et al., 2012). Lower levels (site AL1 and part of site AL2) of the sampled succession, however, yielded reversed directions, and these strata must therefore predate the superchron, but are no older than the base of M0r (126.3 Ma; Gradstein et al., 2012). We thus conclude that the sampled succession spans at least a lowermost to mid-Aptian age range (126–118 Ma).

Figure 3. Rock magnetic and paleomagnetic results from locality AL. (A) Thermomagnetic curve measured on a Curie balance. Heating in black; cooling in gray. (B) Intensity decay curve. (C–F) Orthogonal vector diagrams for in situ coordinates, where closed (open) symbols indicate declination (inclination). (G) Bootstrapped fold test, with cumulative distribution function (with confidence interval in light gray) based on 1000 bootstraps (average of bootstraps in darker gray). (H) Characteristic remanent magnetization (ChRM) values from site AL4, in comparison to the geocentric axial dipole (GAD) field (gray diamond). (I) ChRM values from sites AL1, AL2, AL3, and AL5 in tilt-corrected coordinates. (J) Virtual geomagnetic pole (VGP) positions of sites AL1, AL2, AL3, and AL5.

PALEOMAGNETIC DATABASE FOR GUERRERO

We supplemented our new data on the paleolatitudinal history of the Guerrero terrane with all those previously published that we could find (Fig. 5; Table DR2 [see footnote 1]). In general, papers presenting paleomagnetic data do not supply the original paleomagnetic directions per sample, but only site averages and their statistical parameters (N, k, α_{95}). This is not problematic for lava sites, whereby acquisition of the natural remanent magnetization occurs geologically instantaneously upon cooling of the lava, and the recorded direction is a spot reading of the paleomagnetic field. For sedimentary or (slowly cooling) intrusive rocks, however, each individual sample cannot be considered a spot reading, and it includes a certain amount of PSV. It is therefore preferred to perform statistics on the actual distribution of individual directions rather than on site averages (Deenen et al., 2011). To minimize this bias, we created parametrically sampled data sets for sites of intrusive or sedimentary rocks. For lava sites and (fast cooling) intrusive rocks with $k \geq 50$, we used site averages as single directions.

For the construction of the paleomagnetic database, we selected data according to the following quality criteria: (1) The total distribution of directions satisfies the quality criteria of representing PSV (i.e., $A_{95min} < A_{95} < A_{95max}$; Deenen et al., 2011); (2) sedimentary and intrusive rocks have $n \geq 4$, lavas have $k \geq 50$ (Biggin et al., 2008; Johnson et al., 2008), and at least seven lava (or fast-cooling intrusive) sites could be averaged from a small area; (3) when not corrected for bedding-tilt (in their in situ orientations), directions differ significantly from the local GAD field direction; and (4) bedding-tilt corrections are applied. Sites discarded by the original authors were not taken into account, if reasons were provided.

The database (Fig. 5) contains data sets consisting of 7–14 individual directions for the magnetic field (labeled low quality) and data sets containing ≥20 directions (labeled high quality). High-quality data are only available for the Late Cretaceous and younger history of the Guerrero terrane. The previously published and new poles show that paleolatitudes of the Guerrero terrane do not differ significantly from the paleolatitudinal path predicted for the North American plate, both after collision, but notably also prior to collision (Fig. 5). In other words, the motions of the Guerrero terrane in Early Cretaceous time, when it was separated from the North American plate by a divergent plate boundary, were longitudinal only, and the Guerrero terrane did not undergo significant northward or southward motion relative to the North American plate. Most paleomagnetic declinations indicate counterclockwise rotation, mainly concentrated in the Cretaceous, with perhaps a small rotation occurring after ca. 30 Ma.

DISCUSSION

Paleomagnetic data from the Guerrero terrane yield a paleolatitudinal plate-motion history equal to that of the North American plate since Early Cretaceous time. This implies that, even though the Guerrero terrane must have been located on a separate "Guerrero" plate in Aptian time, relative plate motion between the North American plate and the Guerrero plate was longitudinal only. Our new data are thus consistent with paleogeographic models, based on field and petrologic data, in which the Guerrero arc successions formed on North American crust that rifted away from the Mexican mainland by opening of a back-arc basin above an eastward-dipping subduction zone (Fig. 6), as proposed by, e.g., Cabral-Cano et al. (2000), Elías-Herrera et al. (2000), Martini et al. (2009), and Martini et al. (2014). Additionally, it contradicts paleogeographical models that place the

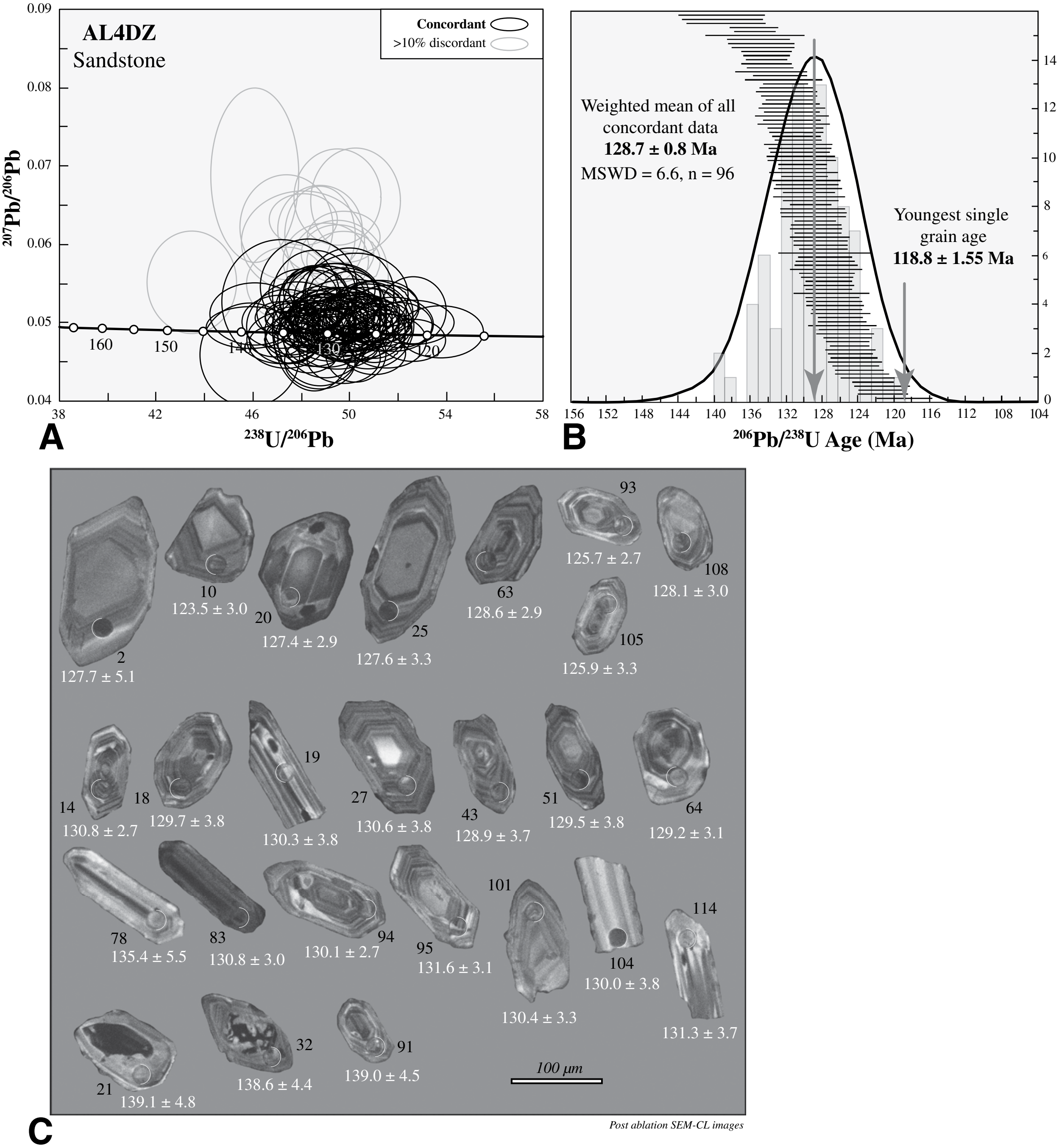


Figure 4. U-Pb age of sample AL2DZ. (A) Concordia diagram for U-Pb ratios. (B) $^{206}Pb/^{238}U$ ages. (C) Postablation scanning electron microscope (SEM) cathodoluminescence (CL) images of selected zircon crystals from this sample. MSWD—mean square of weighted deviation.

Guerrero terrane on the Farallon plate (Dickinson and Lawton, 2001; Umhoefer, 2003; Sigloch and Mihalynuk, 2013), which would predict latitudinal motion between the Guerrero terrane and the North American plate prior to accretion (Doubrovine and Tarduno, 2008; Wessel and Kroenke, 2008; Seton et al., 2012; Wright et al., 2016). Assessing the validity of the paleogeographical model of Hildebrand (2013, 2015) by plate kinematic data is difficult, because this model does not define tectonic plates, nor does it indicate the relative latitudinal plate motions of the Rubian ribbon continent prior to its postulated ca. 125 Ma collision with North America. However, Hildebrand (2013, p. 97) stated, "relative to North America the Rubian ribbon continent migrated south until 80–75 Ma, when its relative displacement reversed to dextral." These postcollisional southward and northward shifts are not in agreement with paleomagnetic data from the Guerrero terrane. Furthermore, the mismatch between the Hildebrand hypothesis and the large body of geological studies suggesting a North American provenance for the

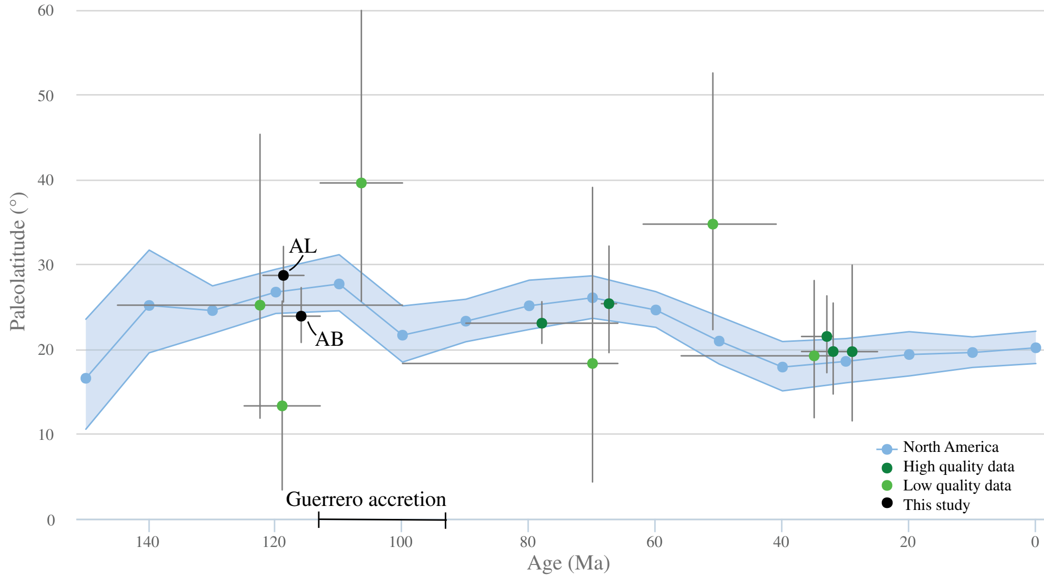


Figure 5. Overview of paleolatitudes of the Guerrero terrane, including data from Böhnel et al. (1989), Goguitchaichvili et al. (2003), Molina Garza et al. (2003), Molina Garza and Ortega Rivera (2006), Rosas-Elguera et al. (2011), Andreani et al. (2014), and this study. The data are plotted with respect to the expected paleolatitudes of the North American plate, calculated for a reference location within the Guerrero terrane (chosen at 20°N, 102°W). We used the reference values predicted by the global apparent polar wander path (GAPWaP) of Torsvik et al. (2012).

Guerrero terrane (e.g., Cabral-Cano et al., 2000; Elías-Herrera et al., 2000; Centeno-García et al., 2008, 2011; Martini et al., 2011) indicates either that the Guerrero terrane was not part of the Rubian ribbon continent, or that this paleogeographic model should not be considered as a plausible model for the assembly of the Mexican Cordillera.

As paleomagnetism cannot constrain paleolongitudes, the magnitude of longitudinal plate motion of the Guerrero terrane relative to North America related to the opening and closure of the Arperos Basin remains unconstrained. During deposition of the Alberca Formation, sediments with a continental provenance did not reach the Guerrero terrane, as evidenced by the absence of any pre-Cretaceous zircons. This may give an indication of a substantial width for the Arperos Basin, although this might also be explained by the presence of significant morphological features within the basin (e.g., the spreading ridge) that prevented mixing of sediments derived from opposite sides. The presence of inherited cores within the zircon crystals suggests that the melts forming the mainly Barremian magmatic rocks intruded older crustal material. An earlier study on the provenance of arc-related volcaniclastic units of the Guerrero terrane (Talavera-Mendoza et al., 2007) demonstrated sediment influx through grain recycling from a variety of ultimately North and South American continental sources.

In recent years, the subduction evolution of the Cordillera has received increased interest because it contains the key to interpreting the complex mantle structure below North America (e.g., Sigloch et al., 2008; Sigloch, 2011; van der Meer et al., 2012, 2017; Sigloch and Mihalynuk, 2013). Tomographic images of the lower mantle below North America, in particular, and the eastern Pacific Ocean reveal large numbers of positive wave speed anomalies that are interpreted as subducted slab remnants (van der Meer et al., 2012, 2017; Sigloch and Mihalynuk, 2013). These studies argued that some (van der Meer et al., 2012, 2017) or all (Sigloch and Mihalynuk, 2013) of these slab remnants resulted from Mesozoic intra-oceanic subduction within the northeastern Panthalassa Ocean. In this context, the Guerrero terrane would be correlated to either the westward-subducted Mezcalera slabs (Sigloch and Mihalynuk, 2013), which correspond to the Hatteras slab of van der Meer et al. (2017), or the equatorial Malpelo slab (van der Meer et al., 2017), both of which were interpreted to be the result of intra-oceanic subduction, suggesting an exotic nature for the Guerrero terrane. Previous field and petrologic studies and our paleomagnetic analysis indicate that formation of the Guerrero terrane was

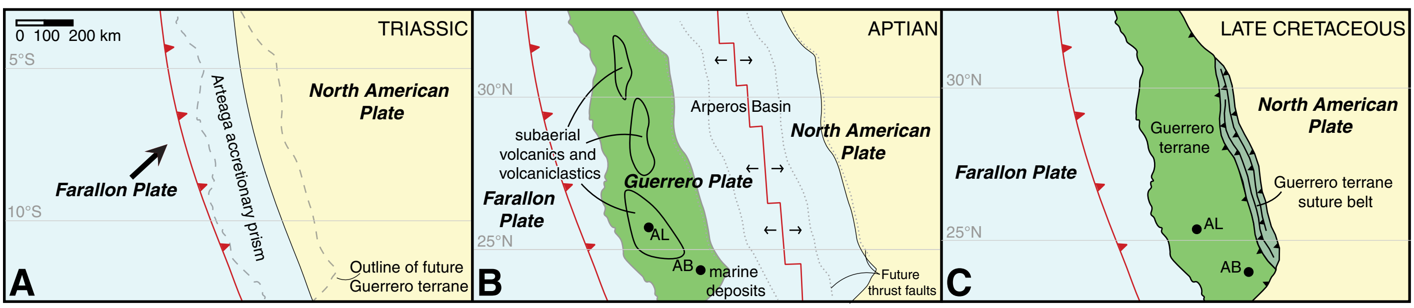


Figure 6. Mesozoic tectonic evolution of the Guerrero terrane. (A) Triassic: continental margin subduction, and development of the Arteaga accretionary prism. (B) Aptian: spreading within the Arperos Basin, formation of the Guerrero plate, and deposition of AB limestones and AL red beds. (C) Late Cretaceous: closure of the Arperos Basin, accretion of the Guerrero terrane to the Mexican mainland, and formation of the Guerrero terrane suture belt.

instead the result of a phase of overriding plate extension within the long-lived eastward continental margin subduction system. A reinterpretation of the correlations between slabs and arcs in the Cordillera is beyond the scope of this paper, but our results necessitate revision of the previous correlations between the Guerrero terrane and lower-mantle slab remnants, and they may provide a strong tie point for the correlation of the western North American continental margin to the mantle in Mesozoic times.

CONCLUSIONS

(1) Paleomagnetic data from the Guerrero terrane yield a paleolatitudinal plate-motion history equal to that of the North American plate since Early Cretaceous time.

(2) This confirms the previously developed paleogeographic model in which the Guerrero arc successions formed on North American crust that rifted away from the Mexican mainland during opening of a back-arc basin above an eastward-dipping subduction zone.

(3) These results contradict alternative paleogeographical models in which the Guerrero terrane is considered to be exotic to the North American continent.

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