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Absolute geomagnetic paleointensity after the Cretaceous Normal Superchron and just prior to the Cretaceous-Tertiary transition

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[1] We report a detailed rock magnetic and Thellier paleointensity study from Autlan volcanic succession in western Mexico. The whole rock K-Ar age obtained for these lavas, 67.4 ± 1.2 Ma, together with the magnetostratigraphy results indicates that the lava flows were erupted during chron 30n just prior to Cretaceous-Tertiary transition. Thirty-nine samples from eight cooling units were preselected for paleointensity experiments on the basis of their low-viscosity index, stable remanent magnetization, and close to reversible continuous thermomagnetic curves. High-quality determinations were obtained for 18 individual samples from four independent lava flows. The mean virtual dipole moment (VDM) obtained in this study is $4.9 \pm 0.6 \times 10^{22}$ A m^2, which is significantly lower than the present geomagnetic field strength but in reasonably good accord with other comparable quality determinations between 35 and 80 Ma. Our data are not significantly different from almost all published paleointensities between 5 and 90 Ma. Moreover, rigorously selected worldwide paleointensity data in the same interval better fit trend 2 of *Heller et al.* [2002]. A careful examination of high-quality paleointensity data from Cretaceous indicates that there is no simple relation between magnitude of Earth's magnetic field and reversal rate. *INDEX TERMS:* 1521 Geomagnetism and Paleomagnetism: Paleointensity; 1510 Geomagnetism and Paleomagnetism: Dynamo theories; 1522 Geomagnetism and Paleomagnetism: Paleomagnetic secular variation; 1540 Geomagnetism and Paleomagnetism: Rock and mineral magnetism; 1560 Geomagnetism and Paleomagnetism: Time variations—secular and long term; *KEYWORDS:* paleointensity, Cretaceous, paleomagnetism

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1. Introduction

[2] The Cretaceous is a key interval in the history of the Earth's magnetic field. The currently debated relationship between the frequency of reversals, secular variation, and paleointensity should be clearly expressed during Cretaceous Normal Superchron (CNS [*Tarduno et al.*, 2001]) when the reversal rate was zero. It has long been argued that low paleointensity prevailed during the Mesozoic [*Prévot et al.*, 1990; *Perrin and Shcherbakov*, 1997]. *Pick and Tauxe* [1993] suggested extension of the Mesozoic dipole low into the whole Cretaceous time. Recently, available reliable paleointensity data [*Goguitchaichvili et al.*, 2002; *Tarduno et al.*, 2001, 2002] suggest, however, that the paleostrength during early Cretaceous may have been comparable or even higher than present intensity and not "anomalously" low as suggested in previous studies. *Riisager et al.* [2002] analyzed worldwide paleointensity data and showed that there are only 15 time-averaged dipole moments between 5 and 160 Ma. This paucity of data makes it difficult to derive any firm conclusions about the evolution of geomagnetic intensity through geological time.

[3] In this study, we contribute to the investigation of the long-term variation of geomagnetic field strength by reporting new reliable paleointensities from ~67 Ma volcanic rocks from western Mexico. These rocks formed during an interval when the reversal rate was increased from zero to 1.08 reversal per million years [*McFadden and Merrill*, 1997]. We also present analyses of the currently available paleointensity data for the last 160 Myr using reliability criteria for paleointensity measurements.

2. Sampling Details and Age Determination

[4] SW Mexico includes several suspect terrenes accreted in Late Mesozoic to early Tertiary time (Figure 1a). The study area is located in the Guerrero terrene, which comprises sedimentary, volcano-sedimentary and volcanic rocks [*Campa and Coney*, 1983; *Monod et al.*, 1993]. Figure 1b is a simplified geologic map of Autlan area showing mainly

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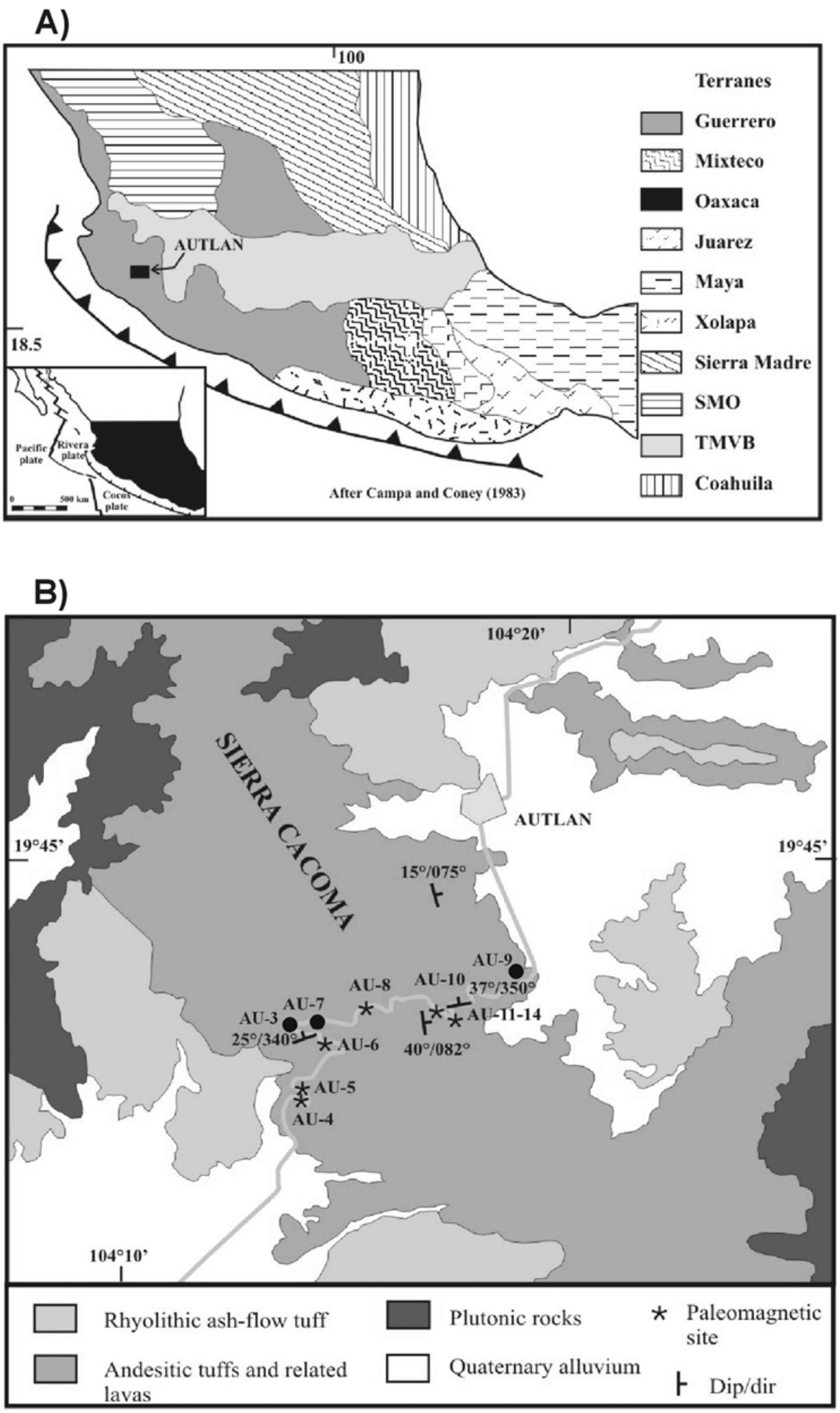

Figure 1. (a) Several suspect terranes accreted in Late Mesozoic to early Tertiary time make up southwestern Mexico [after *Campa and Coney*, 1983]. (b) A simplified geologic map of Autlan area with indication of paleomagnetic sites (see also text).

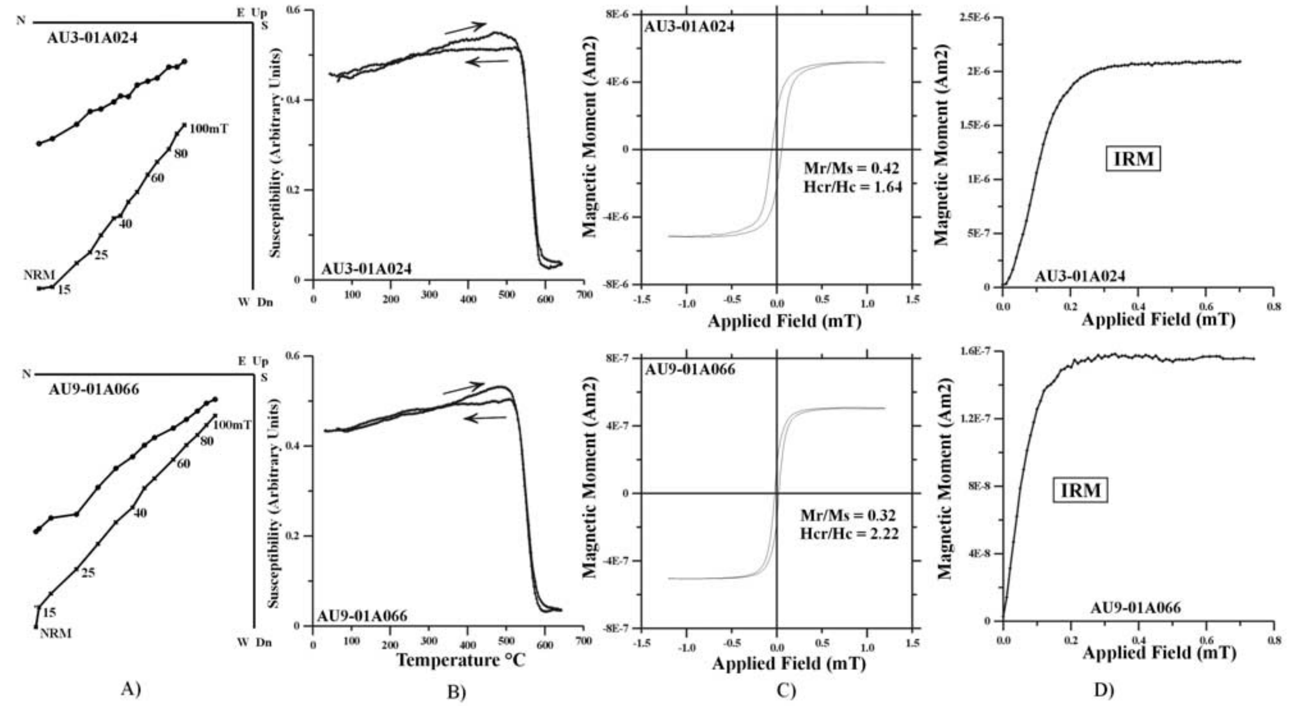

Figure 2. Summary of the magnetic characteristics of the typical samples, selected for Thellier paleointensity experiments. (a) Orthogonal vector plots of stepwise alternating field and thermal demagnetization (stratigraphic coordinates). The numbers refer to peak alternating fields in mT. Circles are projections into the horizontal plane, and crosses are projections into the east-up vertical plane. (b) Initial susceptibility versus temperature curves. The arrows indicate the heating and cooling curves. (c) Examples of hysteresis loops (uncorrected) of small chip samples. (d) IRM acquisition curves.

magmatic units. The intrusive rocks are part of the Puerto Vallarta batholith with ages ranging between 100 and 75 Ma [*Schaaf et al.*, 1995]. The batholith, part of a plutonic belt parallel to the present-day Middle America Trench, intrudes Cretaceous sedimentary and volcano-sedimentary sequences (not shown in the map) forming manganese-rich ore deposits. We sampled a ~300 m thick volcanic succession of younger volcanic rocks that overlie these sequences. The studied units are mainly andesitic basalts with occasionally intercalated ash flow tuffs. A striking feature of the basaltic rocks is the occurrence of large plagioclase phenocrysts (up to 2 cm).

[5] In total, about 150 oriented samples from 15 volcanic units were collected at two separate vertical transects (Figure 1b). Commonly, the outcrops extend laterally over a few tens of meters, and in these cases we typically drilled 7–12 standard paleomagnetic cores per site. The samples were distributed throughout each flow both horizontally and vertically in order to minimize effects of possible block tilting and lightning. Cores were obtained with a gasoline-powered portable drill and then oriented with both magnetic and Sun compasses.

[6] The sample for K-Ar geochronology was taken from site AU3 (middle part of sequence) and prepared by crushing several kilograms of rock, sieving, washing, and separating with magnetic separator and heavy liquids to eliminate the magnetic fraction and leave a feldspar concentrate. The analyses were made at Geology Institute of the National University of Mexico with a VG 1200 noble gas mass spectrometer following the isotope dilution technique. Analytical precision on argon measurements was better than 0.5%. Potassium was determined by X-ray fluorescence analysis following the method described by *Solé and Enrique* [2001] and a 1% relative error is assumed. The age was calibrated with the international standards LP-6 and HD-B1 biotites. The constants recommended by *Steiger and Jäger* [1977] were used in all calculations. The absolute age is 67.4 ± 1.2 Ma (1.99% K, 85.8% $^{40}Ar^*$).

3. Rock Magnetic Properties of Selected Samples

[7] Although hysteresis measurements combined with isothermal remanent magnetization (IRM) experiments were performed on all samples, preselection of the samples for Thellier paleointensity experiments was mainly based on the demagnetization of natural remanent magnetization and temperature dependence of initial magnetic susceptibility. Magnetic characteristics of typical samples selected for Thellier paleointensity experiments are summarized in Figure 2 and are as follows:

[8] 1. Samples are likely to have a small Brunhes age viscous remanent magnetization. In storage tests [*Prévot et al.*, 1983], most samples exhibited viscosity indexes less than 6%.

[9] 2. Selected samples carry essentially a stable, univectorial remanent magnetization, observed upon alternating field treatment (Figure 2a). The median destructive fields range mostly in the 40–50 mT interval, suggesting

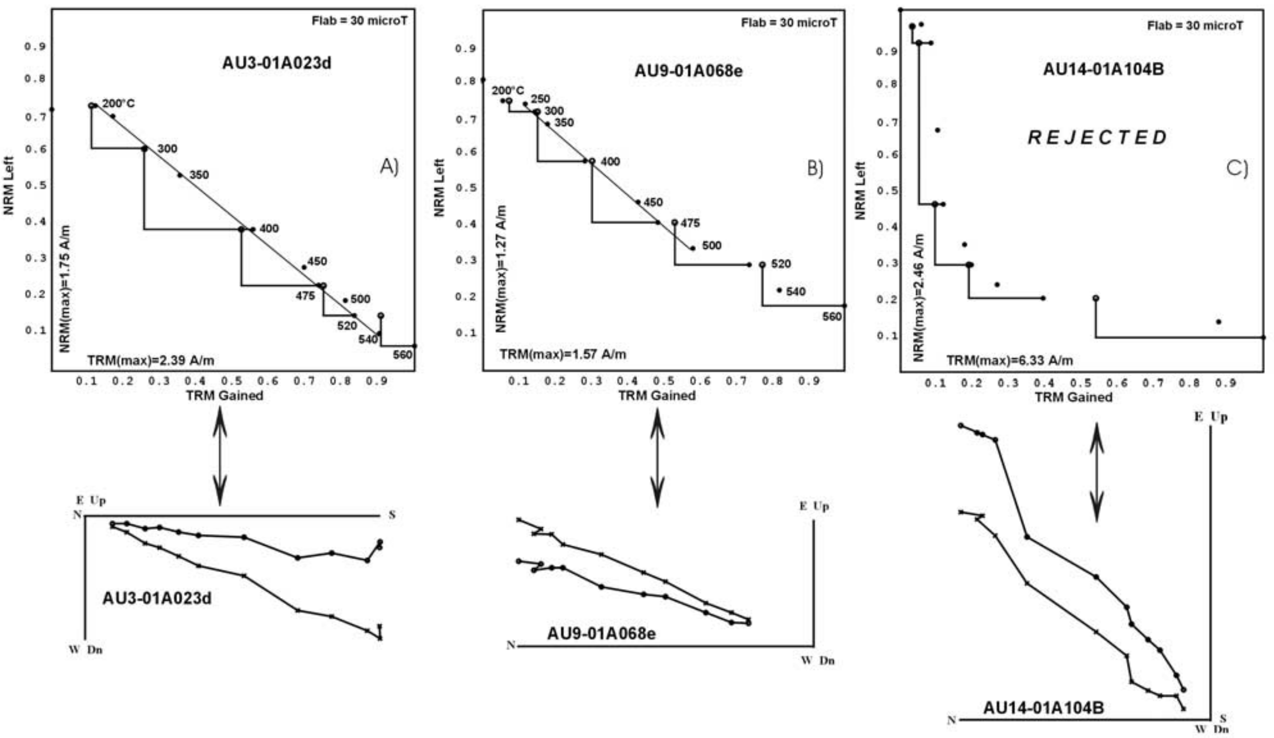

Figure 3. Representative NRM-TRM plots and associated orthogonal vector demagnetization diagrams for Autlan samples. In the orthogonal diagrams we used same notations as in Figure 2a.

the existence of "small pseudosingle domain grains" as remanence carriers [*Dunlop and Özdemir*, 1997].

[10] 3. Low-field continuous susceptibility measurements performed in air (using a Bartington susceptibility meter MS2 equipped with furnace) show the presence of a single ferrimagnetic phase with Curie temperature compatible with Ti-poor titanomagnetite (Figure 2b).

[11] 4. Hysteresis measurements at room temperature show (Figure 2c) that the studied samples fall in the "small pseudosingle-domain" grain size region on a plot Mr/Mrs versus Hcr/Hc [*Day et al.*, 1977]. This probably indicates a mixture of multidomain and a significant amount of single-domain (SD) grains [*Parry*, 1982]. IRM acquisition curves show (Figure 2d) the saturation at moderate fields (150–200 mT), which point to the presence of titanomagnetite. We note, however, that room temperature hysteresis parameters have limited resolution in estimating domain state of most natural rocks [*Goguitchaichvili et al.*, 2001a]. In total, we selected 39 samples from 8 lava flows for the paleointensity experiments having the magnetic characteristics described above.

4. Thellier Experiments

[12] Paleointensity experiments were performed using the Thellier method in its classic form [*Thellier and Thellier*, 1959]. All heatings were made in vacuum better than 10^{-2} mbar. Eleven temperature steps (Figure 3) were distributed between room temperature and 560°C, and the laboratory field was set to 30 μT. Control heatings, commonly referred to as partial thermoremanent magnetization (pTRM) checks [*Prévot et al.*, 1985], were performed after every second heating step throughout the whole experiment. All remanences were measured using both JR5A and JR6 spinner magnetometers.

[13] We accepted only determinations that satisfied all of the following requirements:

[14] 1. Determination obtained from at least six normal remanent magnetization-thermal remanent magnetization (NRM-TRM) points corresponding to a NRM fraction, f [*Coe et al.*, 1978], larger than about 1/3 with quality factor, q, of about 5 or more (Table 1). In two cases (samples 01A048D and 01A054C, Table 1), we accepted determinations with slightly lower q factors because the intensity was found close to the site-mean value and f was 0.35 and 0.46, respectively (Table 1).

[15] 2. There are at least three positive pTRM checks. We define pTRM checks as positive if the repeat pTRM value agrees with the first measurement within 10%.

[16] 3. The directions of NRM end points at each step obtained from paleointensity experiments are stable and linear pointing to the origin. No significant deviation of NRM remaining directions toward the direction of applied laboratory field was observed.

[17] 4. For accepted determinations, γ values (the ratio of potential chemical remanent magnetization (CRM)(T) to the magnitude of NRM(T) for each double heating step in the direction of the laboratory field during heating [*Goguitchaichvili et al.*, 1999a]) are <10° which attest that no significant crystalline remanent magnetization (CRM) is acquired during the laboratory heating.

[18] Eighteen samples, from four individual lava flows, yielded acceptable paleointensity estimates (Figure 3 and Table 1). For these samples the NRM fraction f used for

Table 1. Paleodirectional and Paleointensity Results[a]

Site	n/N	D, deg	I, deg	a_{95}, deg	k	Sample	N'	$T_{min} - T_{max}$, °C	f	g	q	$F_E \pm \sigma(F_E)$, μT	VDM, 10^{22} A m^2	VDMe, 10^{22} A m^2
AU3	8/8	343.2	43.3	5.1	122	01A018D	8	250–520	0.67	0.86	11.4	20.4 ± 1.2	4.24	4.9 ± 0.5
						01A020E	7	250–500	0.47	0.82	8.2	27.1. ± 1.3	5.64	
						01A022B	10	200–540	0.76	0.86	22.1	23.4 ± 0.7	4.87	
						01A023D	10	200–540	0.78	0.87	26.3	24.1 ± 0.6	5.01	
						01A024B	9	200–520	0.70	0.84	20.2	24.8 ± 0.8	5.16	
						01A025D	9	250–540	0.72	0.84	10.2	22.6 ± 1.3	4.70	
AU7	8/8	346.1	49.1	6.8	119	01A045A	9	200–520	0.51	0.83	6.2	25.7 ± 1.7	5.02	5.4 ± 0.7
						01A048D	8	200–500	0.35	0.78	4.3	23.9 ± 2.0	4.67	
						01A049E	7	300–540	0.61	0.82	11.7	29.2 ± 1.1	5.71	
						01A052E	7	300–540	0.65	0.81	5.6	31.7 ± 1.8	6.20	
AU8	6/7	336.3	50.7	8.5	63	01A054C	7	250–500	0.46	0.78	3.6	46.8 ± 3.8	8.98	8.5 ± 0.7
						01A060C	7	200–475	0.33	0.80	6.2	41.7 ± 4.0	8.00	
AU9	8/8	338.4	39.5	7.5	55	01A062C	10	200–540	0.59	0.86	7.5	17.2 ± 1.0	3.71	4.3 ± 0.9
						01A063C	8	250–520	0.65	0.81	8.4	21.7 ± 1.4	4.68	
						01A065D	8	200–500	0.38	0.79	6.3	26.1 ± 1.4	5.63	
						01A066E	8	300–540	0.71	0.82	8.8	19.2 ± 1.0	4.14	
						01A067E	8	300–540	0.83	0.81	11.4	13.9 ± 0.8	3.01	
						01E068E	7	250–500	0.58	0.79	6.5	21.5 ± 1.2	4.64	

[a]Paleointensity results from Autlan volcanic units, n/N is the number of used/treated samples; I and D are magnetic inclination and declination of cleaned remanence of individual samples; k and a_{95} are the precision parameter and confidence radius of 95% on mean; N' is number of NRM-TRM points used for palaeointensity determination; $T_{min}-T_{max}$ is the temperature interval used; f, g, and q are the fraction of extrapolated NRM used, the gap factor, and the quality factor [*Coe et al.*, 1978], respectively. F_E is paleointensity estimate for individual specimen, and $\sigma(F_E)$ is its standard error; VDM and VDMe are individual and average virtual dipole moments, respectively.

determination ranges between 0.33 and 0.83, and the quality factor q varies from 26.3 to 3.6 (generally more than 5). The principal reasons for rejecting paleointensity determination were typical ‘‘concave-up’’ behaviors [*Levi*, 1977; *Kosterov and Prévot*, 1998] (Figure 3c) probably indicating that a substantial part of the sample's TRM is carried by multi-domain grains.

5. Main Results and Discussion

[19] The characteristic paleomagnetic directions determined for Autlan lavas are considered to be of primary origin [*Goguitchaichvili et al.*, 2003a]. This is supported by the occurrence of almost antipodal normal and reversed polarities (Figure 4a). In addition, thermomagnetic investigations show that the remanence is carried in most cases by Ti-poor titanomagnetite, resulting of oxi-exsolution of original titanomagnetite during the initial flow cooling, which most probably indicates thermoremanent origin of a primary magnetization. Moreover, unblocking temperature spectra and relatively high coercivity point to small pseudosingle-domain magnetic structure grains as responsible for remanent magnetization. Single-component, linear demagnetization plots were observed in most cases. Reliable paleomagnetic directions were obtained for 14 sites, corresponding to two normal and twelve reverse polarity

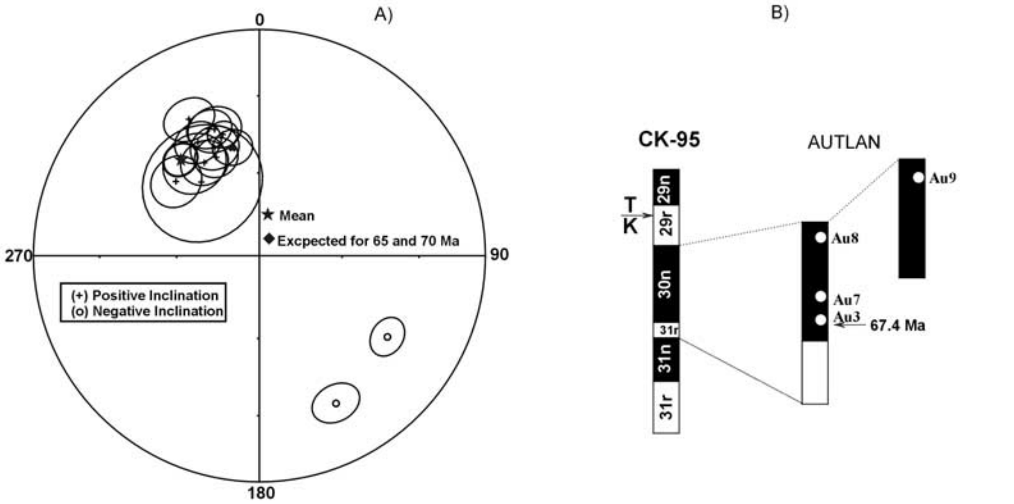

Figure 4. (a) Equal-area projections of the site-mean characteristic paleodirections for the Autlan volcanics (redrawn from *Goguitchaichvili et al.* [2003a]). (b) A tentative magnetostratigraphic correlation between Autlan volcanic units and reference geomagnetic polarity timescale [*Cande and Kent*, 1995]. The relative stratigraphic position of the sites which yielded reliable paleointensity estimates is also shown.

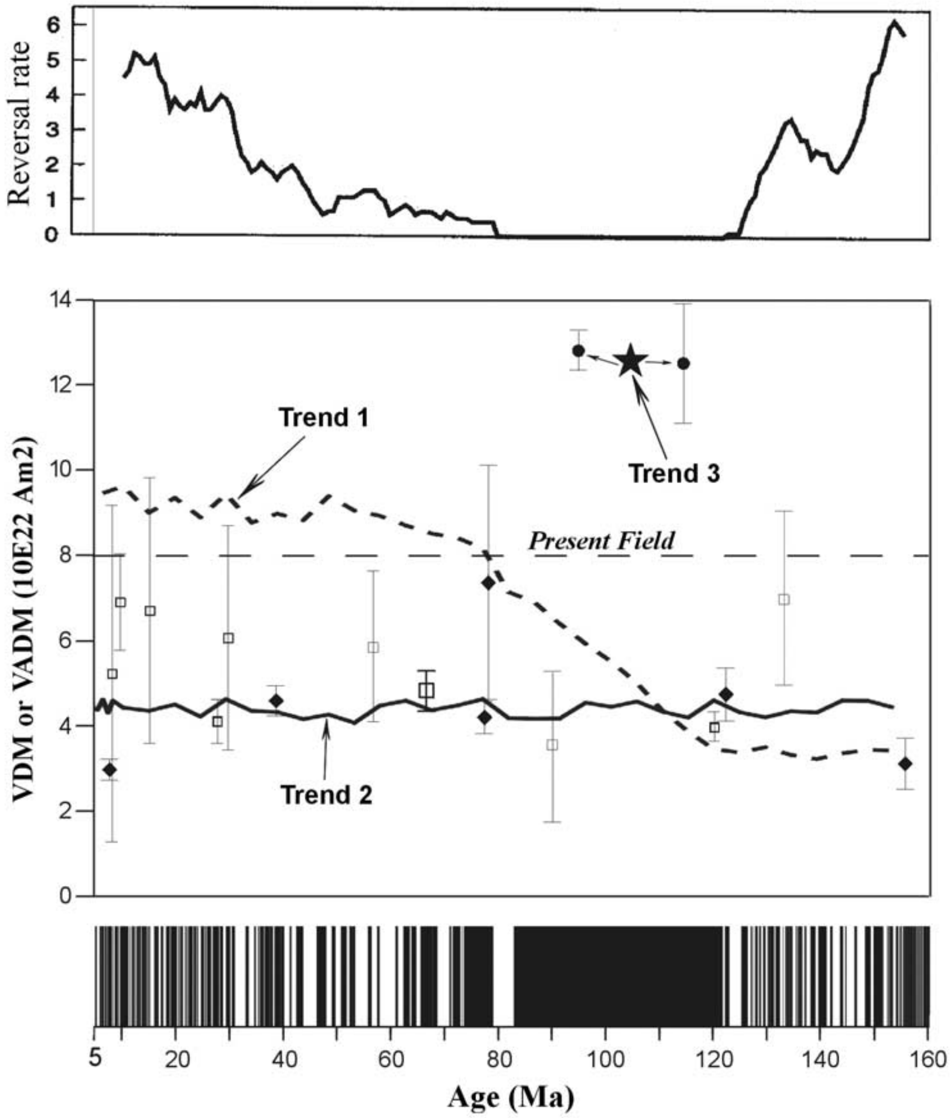

Figure 5. Evolution of mean virtual dipole moments (VDMs) and virtual axial dipole moments (VADMs) for 5 to 160 Ma. Open squares are mean VDM obtained from continental lava flows [*Goguitchaichvili et al.*, 2000, 2001b; *Prévot et al.*, 1985; *Riisager et al.*, 1999, 2000, 2002; *Zhu et al.*, 2001] (this study, large symbols). Diamonds are mean VADMs obtained from submarine basaltic glass [*Selkin and Tauxe*, 2000]. Solid circle is mean VDM obtained for single plagioclase crystals [*Tarduno et al.*, 2001]. Also shown is the geomagnetic polarity timescale to 160 Ma [*Cande and Kent*, 1995] and estimated reversal rate (based on a 10 Myr sliding windows. Short dashed line, solid line, and star show alternative trends for long-term geomagnetic intensity [*Heller et al.*, 2002].

sites [*Goguitchaichvili et al.*, 2003a]. Lowermost flows in the sequence yield reverse polarity magnetization, and probably formed during chron 31r of the reference geomagnetic polarity timescale (Figure 4b). The remaining flows, including those belonging to radiometrically dated site, correspond to chron 30n. These tentative magnetic correlations suggest that the entire volcanic sequence was emplaced during a time span of about 2 Ma.

[20] The mean paleodirection obtained from 14 sites is $I = 44.2°$, $D = 320.6°$, $k = 45$, $\alpha_{95} = 6.0°$. The classical formula $S_F^2 = S_T^2 - S_W^2/n$ was used for estimating paleosecular variation in this study. Here, S_T is total angular dispersion [*Cox*, 1969]

$$S_T = \left[(1/N-1)\sum_{i=1}^{N}\delta i^2\right]^{1/2},$$

N is number of sites used in calculation, δi the angular distance of the ith virtual geomagnetic pole (VGP) from the axial dipole. S_W represents the within site dispersion, and n is average number of samples per unit. *McFadden et al.* [1991] report a VGP scatter equal to about 11.5 for the 19–20° latitude bands between 50 and 80 Ma. We obtained $S_F = 12.1$ (upper and lower limits of 16.1–9.6). Thus Autlan, as a whole sequence, may give time-averaged estimates of mean paleodirections and paleointensity.

[21] The site-mean paleointensity values obtained in this study range from 44.3 ± 3.6 to 19.9 ± 4.2 μT and the corresponding VDMs range from 8.5 ± 0.7 to 4.3 ± 0.9 (10^{22} A m^2). These data yield a mean value of 5.8 ± 1.9 × 10^{22} A m^2. Although both determinations from unit AU8 (Table 1) are of good technical quality, we are forced to reject these data to calculate the mean VDM in order to strictly follow the selection criteria imposed in this study [see also *Riisager et al.*, 2002]. Three flows yielded very similar values giving the mean VDM of 4.9 ± 0.6 × 10^{22} A m^2. Three spot readings of the field strength are probably not enough to represent time-averaged dipole moment even if ideally distributed in 2 Myr interval.

[22] The mean VDM obtained in this study is about 63% of the present geomagnetic axial dipole (7.8 × 10^{22} A m^2 [after *Barton et al.*, 1996]) but in reasonably good agreement with other comparable quality determinations between 35 and 80 Ma. Our data are not significantly different from almost all published paleointensities between 5 and 90 Ma (Figure 5). Long intervals of constant polarity (superchrons) occurred during the Permo-Triassic and the Cretaceous have long been studied to understand the geodynamo and to investigate relationships among polarity reversals, intensity, secular variation, excursions, and morphology of the field. There seems that a general agreement is being reached concerning the direct relationship between long-term variation of the geomagnetic field strength and global geodynamic processes [e.g., *Courtillot and Besse*, 1987; *Larson and Olson*, 1991; *Glatzmaier et al.*, 1999]. *Biggin and Thomas* [2003], for example, propose a model which predicts that changes in paleointensity results from a chain of geodynamic processes extending from crust to core, beginning with plate reorganizations at the surface and culminating in increases in the vigor of outer core convention. One key question is whether changing core-mantle boundary conditions occurring over tens of millions of years are manifested in paleointensity data [*Merrill and McFadden*, 1994; *Heller et al.*, 2002]. Although the paleointensity database has grown significantly over last 10 years, high-quality paleointensity determinations are still scarce and restricted to a few geological locations. This problem is largely due to experimental difficulties related to absolute geomagnetic intensity determination [e.g., *Calvo et al.*, 2002].

[23] Three proposed trends of the long-term paleointensity variation were recently analyzed by *Heller et al.* [2002] (Figure 5). Trend 1 suggests that "average" field strength in the Mesozoic was lower than today (Mesozoic dipole low [*Prévot et al.*, 1990]). Trend 2, which is mainly based on submarine basaltic glass data [*Selkin and Tauxe*, 2000], postulates the relatively high paleointensity since about 0.3 Ma (Figure 5) with lower paleointensities prior to that including whole CNS. In contrast, *Tarduno et al.* [2001, 2002] suggest that the paleointensity was higher during CNS (Trend 3). Figure 5 shows a summary of selected worldwide reliable paleointensity data, including our results for 67 Ma.

[24] It cannot be ascertained that remanence carried by submarine basaltic glass (SBG) is of thermoremanent origin. The excellent technical quality of paleointensity data obtained from basaltic glasses, first underlined by *Pick and Tauxe* [1993] and later by *Juarez et al.* [1998] and *Juarez and Tauxe* [2000], is not by itself proof of geomagnetic fidelity of the paleofield strength found. *Goguitchaichvili et al.* [1999b] showed that the magnetite found in this kind of material may be not of magmatic origin, but crystallizes at rather low temperatures as a result of either glass alteration. In such case, the Thellier paleointensity results would underestimate the actual paleofield strength. Moreover, we still do not have any strict magnetic criteria to distinguish between thermoremanent magnetization (TRM) and crystalline remanent magnetization (CRM). Although, *McClelland* [1996] theoretically predicted that CRM should yield a concave up Arai-Nagata diagrams, *Körner et al.* [1998] and *Goguitchaichvili et al.* [1999b] do not observe this concavity. Microscopic observations may give sometime decisive constraints to attest the TRM, but magnetic grains contained in SBG are too small to be observed directly by the reflected light microscopy. Keeping this in mind, *Goguitchaichvili et al.* [2003b] recently analyzed the paleointensity results from Mio-Plio-Quaternary volcanic rocks. The mean VDM was calculated in two different ways: (1) considering SBG results and (2) discarding them. In both cases, however, quite similar results were obtained. No significant difference was found between the mean (selected) VDMs from SBG and subaerial lavas. Thus we still prefer to keep SBG results on our analyses and in Figure 5.

[25] It can be easily observed that although modest selection criteria [*Riisager et al.*, 2002; *Goguitchaichvili et al.*, 2002] are applied, there are only 17 time-averaged dipole moments between 5 and 160 Ma (Figure 5), which makes difficult to derive any firm conclusions concerning the long-term variation of geomagnetic field strength. We note, however, that these experimental, high-quality data better fit trend 2 of *Heller et al.* [2002] between the 5 and 90 Ma interval. In the 90–160 Ma interval, there are clearly three outliers: two data obtained by *Tarduno et al.* [2001, 2002] and one determination from *Goguitchaichvili et al.* [2002]. Basically, the same conclusion was reached by *Heller et al.* [2002]. The available data indicate no simple apparent relation between the magnitude of Earth's magnetic field and reversal rate (Figure 5).

[26] In summary, the new results, which are correlated to chron 30n (67.4 ± 1.2 Ma), indicate that the preferred mean VDM of 4.9 ± 0.6 × 10^{22} A m^2 is 63% lower than present-day VDM. Therefore it seems that relatively low paleointensities did occur during the interval of increasing polarity reversal rate 83–60 Ma after the CNS. More reliable determinations are strongly needed in particular for the intervals before, during, and after the Cretaceous normal superchron.

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