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A rock-magnetic and paleointensity study of some Mexican volcanic lava flows during the Latest Pleistocene to the Holocene

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Eleven Late Quaternary lava flows were sampled in the Chichinautzin volcanic field of central Mexico to determine their magnetic characteristics and absolute paleointensity. The samples studied cover a geological time interval of approximately 0.39 My to 2000 years. Several rock-magnetic experiments were carried out in order to identify the magnetic carriers and to obtain information about their paleomagnetic stability. Continuous susceptibility measurements with temperature in most cases yield reasonably reversible curves with Curie points close to that of almost pure magnetite, which is compatible with low-Ti titanomagnetite resulting from oxi-exsolution. Judging from the ratios of hysteresis parameters, it seems that all samples fall within the pseudo-single domain grain size region, probably indicating a mixture of multidomain and a significant amount of single domain grains. Forty-two samples belonging to six independent cooling units yielded acceptable absolute paleointensity estimates. The NRM fractions used for paleointensity determination range from 0.34 to 0.97 and the quality factors varies between 4.5 and 97.8, being normally greater than 5. The obtained virtual dipole moment values are higher than those recently reported for the past 5 My and to the present day geomagnetic field strength. Individual paleointensity of around 2000 BP is substantially higher than the present day intensity, which is in broad agreement with worldwide archeomagnetic results.

1. Introduction

The study of absolute paleointensity variation can provide powerful constraints of the physics of the Earth's deep interior and the functioning of the geodynamo. Glatzmaier *et al.* (1999) and Coe *et al.* (2000) recently suggested that absolute intensity should be a fundamental constraint in the recent numerical models that promise to provide unprecedented insight into the operation of the geodynamo. Despite about 40 years of research, paleointensity data are scarce (Perrin and Shcherbakov, 1997) and cannot be yet used to document a long-term variation in the intensity of the Earth's magnetic field through geological time. The major reason for the small number of determinations is that paleointensity is the most difficult component of the magnetic field to determine and the failure rate is often large, generally in the order of 80% (Kosterov and Prévot, 1998).

Prévot *et al.* (1990) first underlined the existence of a relatively low field during the Mesozoic time. However, little is known about the transition mode between the Mesozoic low and present-day 'high' field, largely because of the limited amount of reliable data available. Based on a paleointensity study of submarine basaltic glass, Juàrez and Tauxe (2000) argued that the geomagnetic field strength may have been also substantially low during the Plio-Quaternary time (from 5 to 0.3 My and maybe for the last 160 My), which is in disagreement with the value of approximately 8.2×10^{22} Am^2 often quoted for the last 5 My (see, i.e., Goguitchaichvili *et al.*, 1999). Although numerous absolute intensity determinations are available for the last 50 ky, little is known about the paleostrength of the geomagnetic field during this period, mainly because of unusually high dispersion (Perrin and Shcherbakov, 1997; Perrin *et al.*, 1998). This high dispersion of paleointensity results may be due to from some experimental artifacts and not from the geomagnetic field. In this paper, we report a detailed rock-magnetic and paleointensity study for central Mexican Latest Pleistocene to Holocene volcanic lava flows. The principal objective of this study was to obtain a reliable absolute intensity for the last 50 ky and also to contribute to the archeomagnetism of Mesoamerica, since some of the units studied erupted during this historic time.

2. Sampling Details

The volcanic lava flows selected for the study are located in the southern part of the Basin of Mexico. All sites belong to the *Sierra Chichinautzin* group (Fig. 1), which represent the youngest eruptive activity in the Trans Mexican Volcanic Belt. The eruptive phase of this group probably occurred between the Pleistocene and Holocene times (Urrutia-Fucugauchi and Martin del Pozzo, 1993). These monogenetic volcanoes which outcrop around Mexico City, seem to originate from the subduction of the Cocos Plate under southern Mexico. Volcanic activity developed during the Brunhes geomagnetic period and is likely younger than 40 Kyr in the northern Chichinautzin Sierra (Urrutia-Fucugauchi, 1996; Urrutia-Fucugauchi and Martin del Pozzo, 1993; Gonzalez-Huesca, 1992; Gonzalez-Huesca *et*

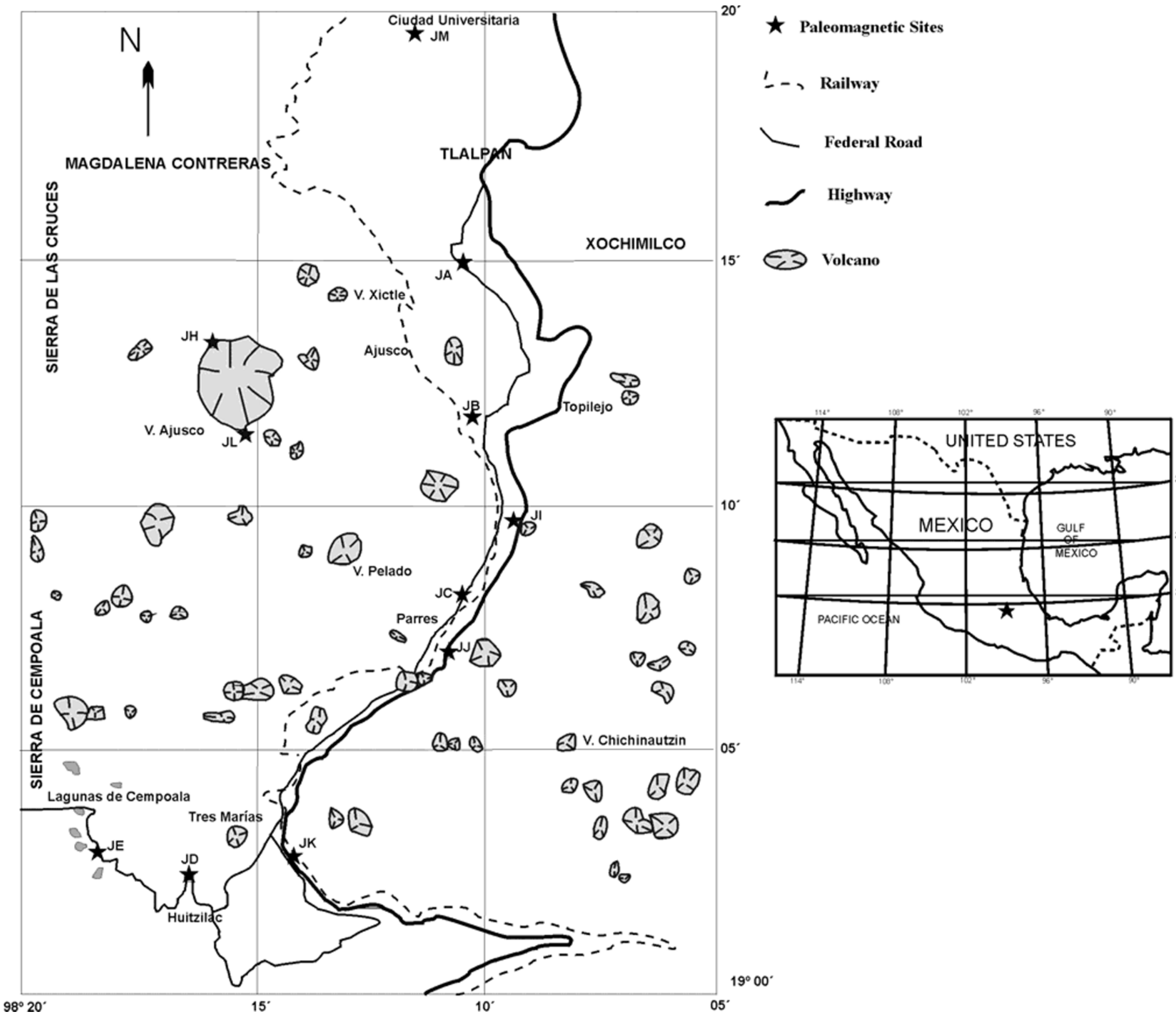

Fig. 1. Schematic map showing the locations of the paleomagnetic sites (stars).

Table 1. Averaged paleodirection of cleaned remanence and corresponding virtual geomagnetic pole (VGP) positions for central Mexican Late Quaternary volcanics. Lat. N and Lon. W are the geographic latitude (north) and longitude (west) of the studied sites. N, number of specimens used for calculation; Dec, declination; Inc, inclination. k and α_{95}: precision parameter and radius of 95% confidence cone of Fisher statistics, Plat/Plong: latitude/longitude of the VGP position, Age: radiometric ages as reported in (1) Gonzalez-Huesca *et al.* (1997), (2) Urrutia-Fucugauchi and Martin del Pozzo (1993), (3) Mora-Alvarez *et al.* (1991), (4) Böhnel *et al.* (1997), and (5) Urrutia-Fucugauchi (1996), Method: radiometric method used for the dating. *In some cases, we demagnetized two samples per core (one by alternating fields and the other one by temperatures) but best quality demagnetization was used for paleomagnetic analyses.

Site	Volcan	Lat. N	Lon. W	N	Dec	Inc	k	α_{95}	Plat	Plong	Age (Years BP)	Method	Ref.
JB	Pelado	19°.11′.14″	99°.10′.15″	8	10.4	17.0	198	3.9	75.6	35.2	4070 ± 150	^{14}C	1
JD	Huilote	19°.02′.05″	99°.16′.19″	8	13.8	10.8	353	3.0	70.9	34.4	<10000		2
JE	Huilote	19°.02′.06″	99°.18′.25″	8	4.0	23.1	277	3.3	82.0	51.7	<10000		2
JH	Ajusco	19°.13′.12″	99°.16′.24″	8	342.7	21.5	371	2.9	71.6	148.3	39000 ± 160	K-Ar	3
JJ	Acopiaxco	19°.06′.60″	99°.10′.54″	13	352.8	33.0	498	1.9	83.1	163.7	<39000		3
JL	Ajusco	19°.11′.55″	99°.15′.00.″	10	358.8	45.0	131	4.2	82.4	152.9	18680 ± 120	^{14}C	2
JM	Xitle	19°.19′.19″	99°.11′.22″	13	352.0	36.0	269	2.5	82.4	179.6	2030 ± 60	^{14}C	4, 5

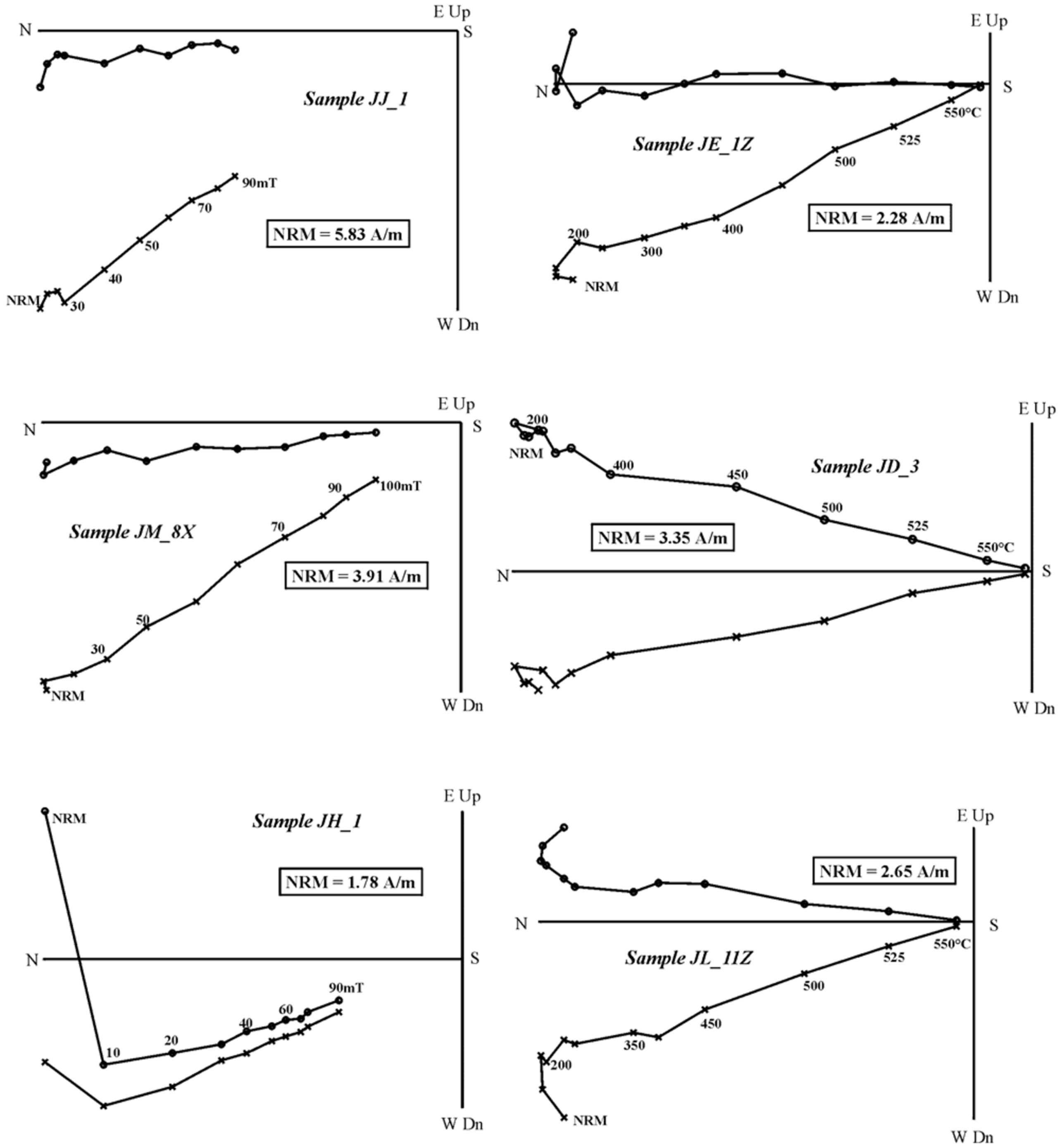

Fig. 2. Orthogonal vector plots of stepwise alternating fields (samples JJ_1, JM_8X and JH_1) and thermal (samples JE_1Z, JD_3, JL_11Z) demagnetization of representative samples (stratigraphic coordinates). The numbers refer to the alternating fields in mT or to temperatures in °C. ○–projections in the horizontal plane, ×–projections in the vertical plane.

al., 1997; Mora-Alvarez *et al.*, 1991). The radiometric dates (using ^{14}C and K-Ar systematics) are available for four out of the seven studied sites (Table 1). Only the relative ages are known for the three remaining volcanic lava flows. The former ages are based on geological observations and previous paleomagnetic studies (Urrutia-Fucugauchi and Martin del Pozzo, 1993). The major-element geochemistry of the lavas was determined by Gonzalez-Huesca (1992) using an atomic absorption spectrometer. When K2O is plotted against the silica content, all the lavas fall into the Cal-calkaline field with basalts and basalt-andesites.

In total, 120 oriented cores belonging to 11 lava flows were collected. Commonly, the outcrops extend laterally over a few tens of meters. In these cases, we drilled 7–15 standard paleomagnetic cores per volcanic unit. The samples were distributed throughout each flow, both horizontally and vertically in order to minimize the effects of block tilting and lightning. Cores were obtained with a gasoline-powdered portable drill, and then in most cases oriented with both magnetic and Sun compasses.

3. Paleodirections

Remanence measurements were made using a Molspin spinner magnetometer (sensitivity $\sim 10^{-8}$ Am^2). For one sample per site, stepwise thermal treatment up to 580°C was carried out using a Schonstedt furnace with a residual magnetic field of about 15 nT. After each heating step, the low field susceptibility was measured at room temperature using a Bartington MS2 susceptibility meter in order to check for possible magnetic changes. For the remaining samples, stepwise alternating field (AF) cleaning was carried out using a laboratory-made AF demagnetizer providing fields of up to 100 mT. Heatings were performed in air.

Natural remanent magnetization (NRM) measurements showed that four of the 11 sampled sites, were characterized by unusually high intensity (more than 100 A/m) and very scattered NRM directions. Both factors may point to a strong lightning-produced magnetization overprint. These samples were rejected for further paleomagnetic analyses. The remaining sites yielded reasonably small within-site NRM direction scatter with α_{95} less than 11°. All 65 samples, belonging to these seven volcanic lava flows, were subjected to detailed magnetic cleaning. Almost all of them carry essentially a single component magnetization, observed upon both alternating field and thermal treatments (Fig. 2). A secondary component, probably of viscous origin, is sometime present and easily removed when applying 10 mT (Fig. 2, sample JH_1) or 200°C (sample JL_11Z). The median destructive fields (MDF) range mostly between 50 and 70 mT, suggesting 'small' pseudo-single domain grains as remanent magnetization carriers (Dunlop and Özdemir, 1997). Some (titano)hematites, together with titanomagnetites, may be present judging from the relatively high coercivity observed. However, we could not detect the presence of (titano)hematites during microscopic observations under reflected light. It is worth noting that the size of the magnetic grains is often too small and the petrological/optical observations of larger ferrimagnetic minerals may not completely correlate with the magnetic behavior of the whole rock. The greater part of the remanence was demagnetized at temperatures of between 500 and 550°C (Fig. 2, samples JE_1Z, JD_3 and JL_11Z), which is common for Ti-poor titanomagnetite resulting from oxy-exsolution. These observations are compatible with thermomagnetic experiments (see below).

The primary magnetization direction was determined by the least-squares method (Kirschvink, 1980), four to nine points being taken for this determination. In some cases, we demagnetized two samples per core (one by alternating fields and the other by temperatures) but best-quality demagnetization was used for paleomagnetic analyses. The obtained directions were averaged by unit and the statistical parameters calculated assuming a Fisherian distribution. The average flow paleodirections are precisely determined (Table 1 and Fig. 3). α_{95} ranges between 1.9° and 4.2°, which corresponds to the high quality paleomagnetic data.

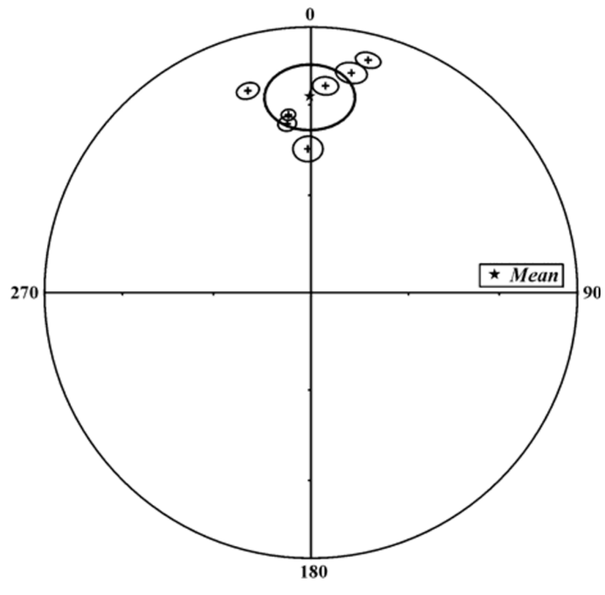

Fig. 3. Equal area projections of the site mean characteristic paleodirections with α_{95} (radius of 95% confidence cone of Fisher statistics).

4. Rock-Magnetic Properties

4.1 Continuous susceptibility measurements

Low-field susceptibility measurements (k-T curves) in the air were carried out to identify the magnetic minerals carrying the remanence and to check their thermal stability. All specimens were heated up to about 600°C at a heating rate of 20°C/min and then cooled at the same rate using a Highmoore susceptibility bridge equipped with a furnace in a Mexico City laboratory. The Curie temperature was determined by the method of Prévot *et al.* (1983). Alternatively, low-temperature (from about −190°C to room temperature) susceptibility was recorded using the same apparatus.

Two different types of behavior were observed during low-T susceptibility experiments (Fig. 4): more than half of the studied samples showed a rather monotonic decrease from about −185°C to room temperature (Site JJ) yielding a very poorly defined inflection point at about −145°C. In the other cases, a rather well-defined pick was observed around −100°C (Fig. 4, Site JH). In both cases, titanium-poor titanomagnetite may be responsible for remanent magnetization. As showed by Özdemir *et al.* (1993), the Verwey transition may be largely suppressed for the titanomagnetites with variable titanium content. Alternatively, similar behavior may also belong to non-stoichiometric (partially oxidized) magnetite. In one case, a quite well defined minimum was detected (site JH) at −60°C. This temperature is substantially different from the Morin transition observed for pure hematite (around −15°, after Dunlop and Özdemir, 1997). Only speculatively can we attribute this minimum to the presence of some (titano)hematites. Let us mention that there is a great deficit in rock-magnetic literature about the interpretation of low-temperature continuous susceptibility curves from natural samples. Two milestone papers (Rhadakrishnamurty *et al.*, 1981; Senanayake *et al.*, 1981) both yield very contradictory results. More recent works (Özdemir *et al.*, 1993, see also Dunlop and Özdemir, 1997) show only results from chemically well-identified synthetic (titano)magnetites.

Corresponding high-T susceptibility experiments (k-T curves) indicate in both cases the presence of a single mag-

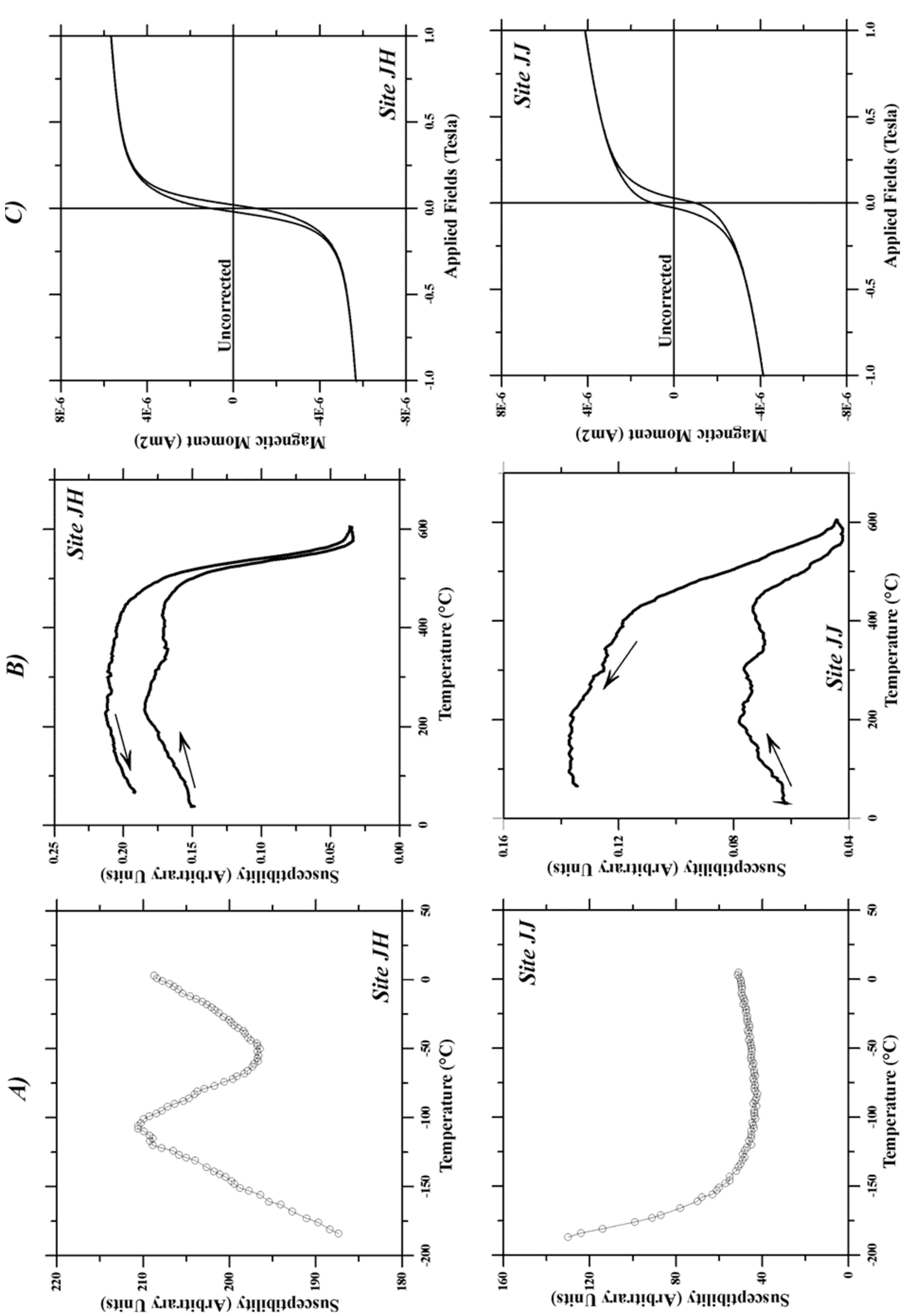

Fig. 4. (A) and (B) Continuous susceptibility versus temperature (low and high temperatures) curves for representative samples. The arrows indicate the heating and cooling curves. (C) Examples of hysteresis loops (uncorrected) of small chip samples (see also the text).

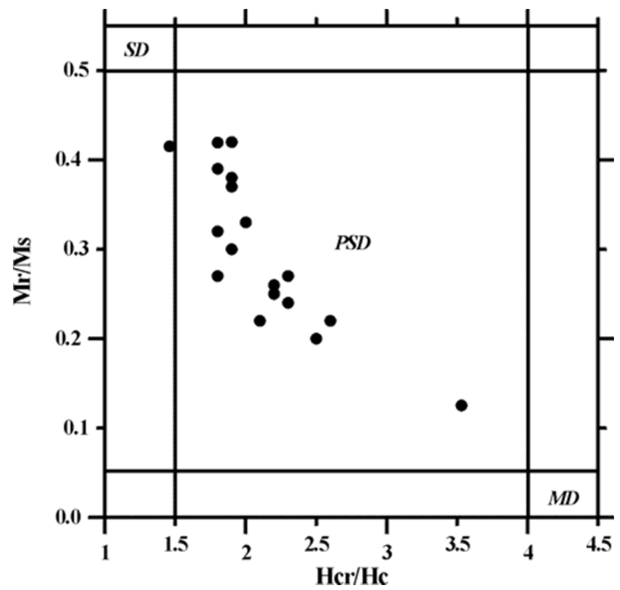

Fig. 5. Day's (1977) diagram showing the possible magnetic domain-type for Mexican Late Quaternary volcanic samples.

netic/ferrimagnetic phase with a Curie point compatible with relatively low-Ti titanomagnetite (around 550°C, Fig. 4). However, the cooling and heating curves are not perfectly reversible because of the relatively low signal of initial magnetic susceptibility. Let us point out here that we were not able to observe the Curie point of (titano)hematite on the k-T curves. Microscopic observations on polished sections also indicate that the main magnetic mineral is probably low-Ti titanomagnetite associated with exsolved ilmenite probably formed as a result of oxidation of titanomagnetite during the initial flow cooling. These intergrowths typically develop higher than 600°C (Haggerty, 1976) and, consequently, the NRM carried by these samples should be a thermoremanent (TRM) magnetization.

4.2 Hysteresis measurements

Hysteresis measurements at room temperature were performed on all studied samples using the AGFM 'Micromag' apparatus of the paleomagnetic laboratory at Mexico City in fields up to 1 T. The saturation remanent magnetization (J_{rs}), the saturation magnetization (J_s), and the coercive force (H_c) were calculated after correction for the paramagnetic contribution. The coercivity of remanence (H_{cr}) was determined by applying a progressively increasing backfield after saturation. Some typical hysteresis plots are reported in the right part of Fig. 4. The curves are symmetrical in all cases. Near the origin, no potbellied and wasp-waisted behavior (Tauxe *et al.*, 1996) was detected, which probably reflects the very restricted ranges of the opaque mineral coercivities. Associated IRM acquisition curves (not shown), which were measured in a 1 T maximum field, were found to be very similar for all samples. Saturation is reached in moderate fields of the order of 150–300 mT, which points to some fine-grained spinels as remanence carriers. J_{rs}/J_s ranges from 0.12 to 0.42 and H_{cr}/H_c varies between 1.4 and 3.6 (Fig. 5). Judging from these values, it seems that all samples fall in the pseudo-single-domain (PSD) grain size region, probably indicating a mixture of multidomain (MD) and a significant amount of single-domain (SD) grains (Day, 1977).

5. Paleointensity Determination

Paleointensity experiments were performed using the Thellier method (Thellier and Thellier, 1959) in its modified form (Coe *et al.*, 1978). The heatings and coolings were made in the air and in the laboratory field was set to 30 or 50 μT. Some 10–11 temperature steps were distributed between 200 and 580°C. Several control heatings (i.e. reinvestigations of results from previous heating steps, commonly referred to as partial TRM (pTRM) checks) were performed throughout the experiments.

Following the paleodirectional and rock-magnetic results, altogether 62 samples, yielding stable magnetization component with relatively high MDF values, elevated blocking temperatures, and reasonably reversible k-T curves, were selected for paleointensity experiments.

Paleointensity data are reported on the Arai-Nagata (Nagata *et al.*, 1963) plots in Fig. 6. We accepted only the following determinations: (1) obtained from at least 5 NRM-TRM points corresponding to a NRM fraction larger than 1/3 (Table 2), (2) yielding a quality factor (Coe *et al.*, 1978) of about 5 or more, and (3) with positive 'pTRM' checks. For these samples, the NRM fraction f used for determining ranges between 0.37 to 0.97 and the quality factor q varies from 4.5 to 97.8 (generally more than 5). Some 42 samples exhibited the above-described behavior and constitute a 'hard core' of the present paleointensity data set. Eight samples yielded apparently linear segments, although lower quality factors, and were discarded in the calculating of mean paleointensity. They are presented in italics in Table 2. No significant changes of magnetic susceptibilities were detected for these samples after each double heating step, which discards any important mineralogical changes during heatings.

6. Discussion and Main Results

We consider the paleomagnetic components determined in this study to be of primary origin for several reasons: (1) thermomagnetic curves show that the remanence is carried in most cases by the 'near magnetite' phase. (2) Reflected light microscopy shows that this (titano)magnetite results from the oxi-exsolution of the original titanomagnetite during the initial flow cooling, which most probably indicates the thermoremanent origin of the primary magnetization. (3) Unblocking temperature spectra and relatively high MDF values point to a small pseudo-single domain low-Ti titanomagnetite as being responsible for remanent magnetization, and (4) single-component, linear demagnetization plots were observed in 75% of cases.

Paleomagnetic results of the present study yield a $D = 359.4°$, $I = 27°$, $k = 28$, $\alpha_{95} = 11.8°$, direction. No significant difference can be recognized with the expected Plio-Quaternary paleodirection (Besse and Courtillot, 1991). Nevertheless, if paleodeclinations of individual sites are regarded, some deviations from the expected direction can be recognized (Fig. 3), which may be related to paleosecular variation.

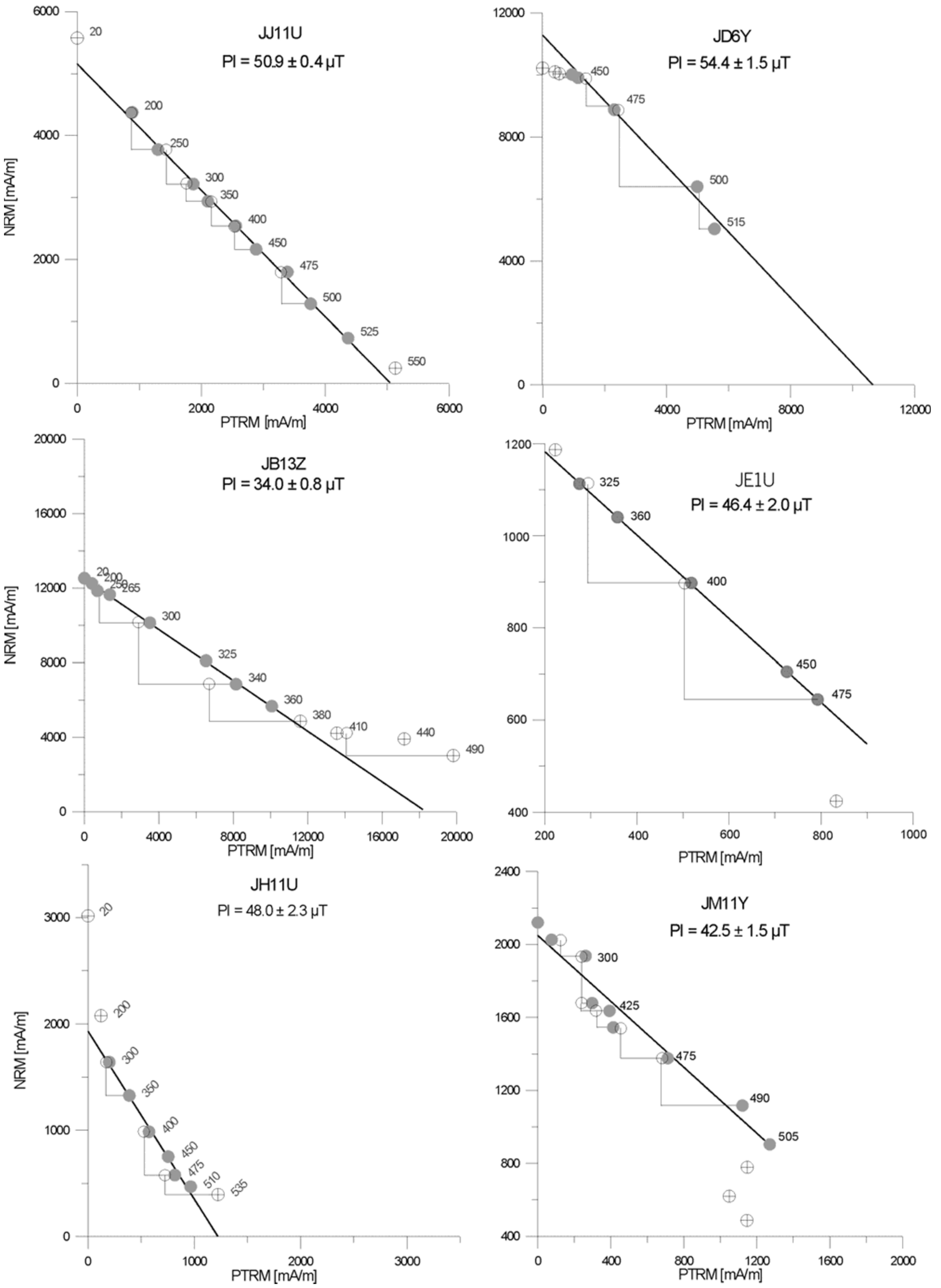

Fig. 6. Representative NRM-TRM (Arai-Nagata) plots for the central Mexican Late Quaternary volcanic lava flows. Open circles refer to the ‘pTRM’ checks. ⊕–Rejected.

Table 2. Absolute paleointensity results from the central Mexican Late Quaternary lava flows. T_{min}-T_{max} is the temperature interval used for paleointensity determination, N is the number of NRM-TRM points used for determination, f, g, and q are the fraction of extrapolated NRM used for intensity determination, the gap factor, and the quality factor (Coe *et al.*, 1978), respectively. PI is the individual paleointensity estimate with associated error. Italic font denotes the lower quality determinations which were discarded in site-mean paleointensity calculation (see the text).

Sample	T_{min}-T_{max}	N	g	f	q	PI (microT)	S.D.
JM5X	200–475	6	0.764	0.477	12.3	55.7	1.7
JM6X	200–500	7	0.808	0.469	8.8	60.5	2.6
JM7X	200–500	7	0.818	0.387	4.5	57.0	4.1
JM9X	*425–525*	*5*	*0.725*	*0.214*	*2.0*	*58.0*	*4.6*
JM10X	20–490	8	0.810	0.606	24.8	53.0	1.1
JM10Y	20–475	7	0.793	0.427	5.6	57.5	3.5
JM11X	200–500	7	0.790	0.569	12.2	44.9	1.7
JM11Y	20–505	9	0.841	0.608	14.5	42.5	1.5
JM12X	200–475	6	0.780	0.604	52.2	55.4	0.5
JM12Y	20–475	7	0.824	0.723	47.6	64.0	0.8
Mean	9 Samples					54.8	6.6
JH10Y	300–510	6	0.772	0.577	20.1	37.9	1.4
JH10Z	300–475	5	0.608	0.614	15.3	34.3	1.4
JH11U	300–510	6	0.773	0.604	16.6	48.0	2.3
JH11Y	300–510	6	0.779	0.635	27.8	47.2	1.4
JH12Y	200–510	7	0.816	0.771	53.8	41.0	0.8
JH12Z	300–510	6	0.774	0.634	13.7	41.9	2.5
JH14X	*450–520*	*5*	*0.723*	*0.467*	*1.7*	*41.0*	*8.0*
JH14Y	*450–520*	*5*	*0.692*	*0.603*	*3.5*	*31.2*	*3.8*
JH17Z	300–510	6	0.751	0.606	55.8	47.8	0.7
Mean	7 Samples					41.1	6.0
JJ2Y	20–450	7	0.795	0.897	54.2	45.6	
JJ10	300–535	8	0.840	0.676	15.7	62.3	2.3
JJ11U	200–525	9	0.868	0.706	77.9	50.9	0.4
JJ11V	200–475	7	0.818	0.554	31.7	45.5	0.7
JJ11X	400–525	5	0.746	0.477	26.8	75.4	1.0
JJ11Y	400–525	5	0.750	0.343	16.1	87.9	1.4
JJ11Z	250–550	9	0.848	0.695	14.0	57.0	2.4
JJ12	250–535	9	0.846	0.754	25.7	66.6	1.7
Mean	8 Samples					61.4	14.9
JB9Z	250–305	5	0.687	0.387	8.1	21.4	0.7
JB9U	250–305	5	0.673	0.509	9.2	20.1	0.8
JB10X	305–450	5	0.667	0.750	4.9	13.4	1.4
JB10Y	*305–450*	*5*	*0.687*	*0.689*	*2.6*	*12.2*	*2.3*
JB12X	290–380	5	0.732	0.370	5.9	27.2	1.3
JB12Y	265–380	7	0.811	0.667	14.3	26.4	1.0
JB13U	20–360	8	0.773	0.608	37.7	34.1	0.4
JB13X	*20–410*	*10*	*0.860*	*0.267*	*4.7*	*36.7*	*1.8*
JB13Y	20–360	8	0.792	0.432	8.7	33.2	1.3
JB13Z	20–360	8	0.795	0.549	19.8	34.0	0.8
Mean	8 Samples					25.9	8.8
JD4X	20–575	7	0.844	0.869	51.3	52.5	0.8
JD4Y	*350–515*	*5*	*0.746*	*0.220*	*1.4*	*68.0*	*8.0*
JD5X	20–575	7	0.771	0.971	97.8	45.8	0.4
JD5Y	*350–500*	*4*	*0.553*	*0.192*	*0.9*	*42.6*	*4.9*
JD6X	20–575	7	0.720	0.892	97.5	53.2	0.4
JD6Y	450–515	4	0.602	0.431	9.7	54.4	1.5
JD7	450–575	5	0.699	0.767	19.6	69.6	1.9
JD8	20–575	7	0.826	0.784	43.6	60.6	0.9
Mean	6 Samples					55.8	9.7
JE1Z	325–525	6	0.745	0.511	9.6	61.7	2.5
JE1U	300–475	6	0.756	0.394	7.1	46.4	2.0
JE2X	*300–475*	*6*	*0.761*	*0.385*	*3.5*	*49.1*	*4.2*
JE2Y	300–475	6	0.781	0.430	5.3	53.9	3.4
JE6Y	350–525	5	0.674	0.444	7.4	85.9	3.5
Mean	4 Samples					59.4	15.9

Table 3. Summary of absolute geomagnetic paleointensities and virtual dipole moments (selected using the same acceptance criteria as in the present study) for Mexican Latest Pleistocene to Holocene lava flows (from Thellier paleointensity experiments only, see also the text). N refers to the number of cooling units used to calculate the average paleointensity (F) and VDM. n refers to the number of individual samples for each cooling unit.

Reference	Age (Years BP)	N	n	$F \pm \sigma_F$ (T)	VDM (10^{22} Am^2)	S.D.
Böhnel *et al.* (1997)	∼2000	1	32	72.1 ± 14.2	16.0	3.2
Gonzalez-Huesca *et al.* (1997)	19000–2000	5	12	48.2 ± 20.6	10.7	4.6
This Study	39000–2000	6	42	49.7 ± 13.6	11.8	2.6
All VDMs		12	86		12.8	2.8

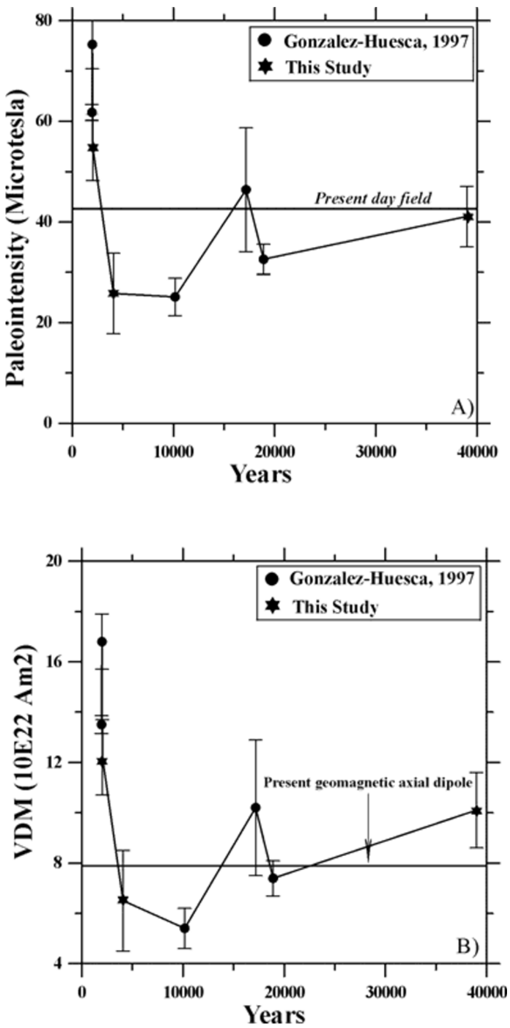

Fig. 7. Evolution of individual paleointensities (a) and virtual dipole moments (b) during the Latest Pleistocene to Holocene time (only results with available radiometric ages are shown) according to selected absolute intensity data from central Mexico (see also the text).

As far as the preliminary palaeointensity experiments are concerned, absolute geomagnetic intensities obtained from Mexican volcanic lava flows vary from 25.9 to 61.4 μT. The mean value is 49.7 ± 13.6 μT. The corresponding mean virtual dipole moment (VDM) is $11.8 \pm 2.6 \times 10^{22}$ Am^2, which is almost double compared to the mean paleointensity for the last 50 ky (6.5 ± 3.3) according to the selected paleointensity data from the IAGA paleointensity database (so-called Montpellier paleointensity database at `ftp://ftp.dstu.univ-montp2.fr/pub/paleointb`, see also Perrin and Shcherbakov, 1997). The average value obtained in the present study, which may be affected by short-term secular variation, is also almost double compared to the mean paleointensity of between 0.3 and 5 My, recently reported by Juàrez and Tauxe (2000) and substantially higher than the present-day geomagnetic field strength (about 8×10^{22} Am^2). Relatively high paleointensities were already reported in Nagata *et al.*'s (1965) pioneering work for some Mexican Quaternary volcanic and archeomagnetic materials. These results have remained essentially unchallenged during the last decade (Gonzalez-Huesca *et al.*, 1997; Urrutia-Fucugauchi, 1996).

Our results, although not numerous, are of high technical quality, as attested by reasonably high Coe *et al.* (1978) quality factors. It is quite hard to assess the reliability of the paleointensity results, due to the poor documentation of many previously published determinations. In our analyses of the Quaternary paleointensity data from Mexico, we consider only results obtained with the Thellier method for which positive pTRM checks attest the absence of alteration during heatings and at least two determinations per flow. Moreover, we imposed the same acceptance criteria on the individual determinations that we use in the present study (i.e. NRM fraction used for a paleointensity determination larger than 1/3 and a quality factor of around or above 5). No transitional data will be used in analyses. From more than 30 absolute paleointensity determinations only five belonging to Gonzalez-Huesca *et al.* (1997) and one from Böhnel *et al.* (1997) fulfill these basic criteria (Table 3, Fig. 7). Combining all reliable paleointensity data currently available from Mexico for the last 50 Ky, we obtain a mean VDM equal to $12.8 \pm 2.6 \times 10^{22}$. It should be noted that there are not enough data to discuss VDM variation through time (Fig. 7) and more determinations are greatly needed.

The individual paleointensity we obtained for about 2000 BP (Fig. 7), is higher than the present field intensity (43 mT). This is in broad agreement with southern U.S.A. (Sternberg, 1989) and worldwide (McElhinny and Senanayake, 1982) archeomagnetic results. Barbetti (1983), however, reported somewhat contradictory results studying southeast Australia aboriginal fireplaces and baked clay. High paleointensity occurrence (some 40% higher than the present field strength) around 2000 BP has been speculated by some authors (Aitken *et al.*, 1989; Böhnel *et al.*, 1997). Our study furnishes evidence fostering that speculation.

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