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DETERMINACIÓN DE LA PALEOINTENSIDAD ABSOLUTA SOBRE LAS ROCAS MEXICANAS Y SUDAMERICANAS: ASPECTOS METODOLÓGICOS E IMPLICACIONES GEOMAGNÉTICAS

T E S I S

q u e p a r a o b t e n e r e l g r a d o d e

D o c t o r e n C i e n c i a s

Presenta

J U A N J U L I O M O R A L E S C O N T R E R A S

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Resumen

La presente investigación esta dedicada, por un lado, a la parte metodológica asociada a la recuperación fiel de los registros de la intensidad del campo geomagnético presentes en las rocas volcánicas, y por el otro, a contribuir de forma confiable a aumentar la base de datos de paleointensidad (PI) global con datos de México y América del sur, incidiendo en las implicaciones geomagnéticas en tiempos geológicos claves para un mejor entendimiento de su evolución temporal.

Así, se revisan las diferentes metodologías desarrolladas para la estimación de la PI y se evalúan las dos mas ampliamente utilizadas para tales fines (Thellier y Thellier, 1959 y Shaw, 1974 y sus variantes). Dado la gran expectación que ha causado el empleo de microondas para la determinación de PI (Shaw, 1993), se aplicó dicha metodología a muestras seleccionadas de flujos históricos del volcán Paricutín para estimar la confiabilidad de esta última alternativa.

Se llevó al cabo una investigación detallada de magnetismo de rocas con la intención de reconocer con anticipación las muestras mas apropiadas para los experimentos de PI, identificando con mayor resolución la mineralogía magnética presente y su estado de dominio magnético empleando criterios termomagnéticos. Se reconoce en la técnica de desmagnetización por temperaturas bajas una herramienta potencialmente útil para aumentar la tasa de éxito de los experimentos de paleointensidades.

En lo que respecta a la parte de las aportaciones a la base global de PI absolutas, se presentan los resultados de una investigación realizada en rocas volcánicas mexicanas miocénicas, para las cuales se obtuvieron los primeros resultados confiables de paleointensidad para Baja California.

Se prestó especial atención a formaciones volcánicas localizadas en sudamérica debido a la gran disparidad en los datos de PI entre ambos hemisferios, evidente de la base de datos de PI global de la IAGA (International Association of Geomagnetism and

Aeronomy), y porque las dos formaciones seleccionadas se encuentran en puntos de periodos clave de la evolución temporal de la intensidad del campo magnético.

El caso de la provincia magmática de Paraná resultó ser especialmente significativo ya que marca un precedente en cuanto a las implicaciones geomagnéticas existentes a la fecha. Los valores de PI promedios por flujo obtenidos en este trabajo dieron como resultado un valor promedio para el Momento Dipolar Virtual (VDM) de alrededor del 90% del valor del dipolo axial presente $(7.2 \pm 2.3) \times 10^{22}$ Am^2 (Barton et al., 1996).

Este resultado, junto con aquellos datos nuevos de Tarduno et al. (2001 y 2002), sugiere una paleointensidad para el periodo Cretácico comparable o mas alta que la actual y no “anormalmente baja” como es sugerido por estudios previos (Prévot et al., 1990). Por tanto, la propuesta del bajo dipolar para el Mesozoico deba ser reconsiderada.

Abstract

The present investigation is aimed to contribute to the methodology of absolute geomagnetic paleointensity determination and increase the global paleointensity database with new reliable data from Mexico and South America. This work discusses about the paleointensity variation during key geologic periods in order to achieve a better understanding of the evolution of the geomagnetic paleostrength through time.

A detailed review of the different methodologies used to estimate absolute paleointensities are made. Moreover, the two apparently most reliable methods (Thellier and Thellier, 1959 and Shaw, 1974) are re-evaluated using Quaternary volcanic rocks from Central Mexico. Due to the great expectation related to the microwave paleointensity determination (Shaw et al., 1996), some selected samples belonging to historic Paricutin volcano were subjected to this treatments in order to estimate the reliability of this alternative technique.

Moreover, a detailed rock magnetic investigation was carried out in order to better recognize the most suitable samples for the Thellier paleointensity experiments with the special emphasis to the identification of the magnetic domain states using thermomagnetic criteria. Low temperature demagnetization is considered as a potentially useful tool in order to increase the success rate in paleointensity experiments.

In regard to the new contributions to the global paleointensity database, the first reliable paleointensity results are obtained from the Miocene volcanic rocks form Baja California.

Special attention was paid to volcanic formations located in South America due to the great disparity in the PI data between both hemispheres, evidenced from the analyses of the global PI database of IAGA (International Association of Geomagnetism and Aeronomy). A detailed rock-magnetic and paleointensity study of Paraná Magmatic Province (PMP), yielded 42 reliable absolute PI determination from 11 independent lava

flows. The average VDM (Virtual Dipole Moment) value obtained for the early Cretaceous is almost 90% of the present field strength. This result, together with the new data of Tarduno et al. (2001 and 2002), suggest a cretaceous PI comparable or even higher that the present day value, and not “abnormally low” as suggested in previous studies.

Introducción

La presente investigación esta dedicada a contribuir a la metodología para la determinación de la paleointensidad (PI) absoluta del campo geomagnético (CGM) y a aumentar la base global de datos de PI con datos confiable de México y Sudamérica. Se discute sobre las variaciones de la PI durante periodos geológicos claves a fin de obtener un mejor entendimiento sobre la evolución de la intensidad del CGM a través del tiempo.

Camps y Prévot (1996) y más recientemente Coe et al., (2000) mostraron que la intensidad absoluta del campo geomagnético es un parámetro decisivo para conocer la morfología del campo geomagnético. Sin embargo, hay pocos datos confiables y no pueden usarse para obtener correctamente las características del campo geomagnético (Riisager et al., 2002). A pesar de más de 30 años de investigación en paleointensidad, la evolución temporal de la intensidad del campo geomagnético es aún poco conocida (Perrin y Scherbakov, 1997; Tanaka et al., 1995). La razón principal para esta incertidumbre es el número limitado de datos confiables disponibles: alrededor de 5 determinaciones por My entre 0-10 My y menos de una por My entre 10 y 400 My. Como consecuencia de esto, solo rasgos a gran escala se pueden definir, como la pronunciada disminución del momento dipolar virtual terrestre (VDM) en el Mesozoico (Prévot et al., 1990), período durante el cual la estructura del campo estuvo preservada (Perrin y Shcherbakov, 1997) al menos entre 260 y 120 Ma. Sin embargo, el término exacto de este período y el modo de transición entre los campos bajo (Mesozoico) y alto (Neógeno) no es conocido, principalmente a causa del número limitado de datos confiables disponibles para esos periodos. Por tanto es necesario obtener más datos para estas edades clave.

En este trabajo se estudiarán algunas formaciones volcánicas de México y América del Sur que presentan varias ventajas: (1) están ampliamente distribuidas en grandes provincias volcánicas y son de fácil acceso; (2) registran fielmente el campo magnético que existió en el tiempo de erupción; (3) son rocas frescas para fechamiento isotópico por K-Ar.

En un intento por aumentar la confiabilidad y el número de este tipo de determinaciones, a lo largo del tiempo se han diseñado diferentes métodos para la determinación de paleointensidades. En la **parte I** de este trabajo se presentan y describen brevemente los diferentes métodos propuestos para tales fines, desde los métodos pioneros hasta los últimos avances al respecto.

Una identificación precisa de los portadores de la magnetización, así como de su estado de dominio magnético, es de fundamental importancia para la determinación de paleointensidades, ya que 1) el significado geomagnético de una determinación depende de que la magnetización natural sea original y sea proporcional a la intensidad del campo magnético al momento de su formación y 2) solo la remanencia portada por granos de dominio sencillo (SD) obedecen las leyes de las magnetizaciones termorremanentes parciales de Thellier, premisa fundamental para la aplicación del método de paleointensidades que lleva su nombre (Thellier & Thellier, 1959). Por lo anteriormente expuesto, se planeó y llevó al cabo una serie de experimentos de magnetismo de rocas sobre especimenes bien controlados para hacer una comparación entre los tres tipos de curvas termomagnéticas comúnmente más utilizadas en paleomagnetismo y magnetismo de rocas. Los resultados preliminares de dicha investigación se presentaron en la "Reunión anual de la Unión Geofísica Mexicana" del año 2000, y como resultado de estos estudios surgieron los artículos *"On the Use of continuos thermomagnetic curves in paleomagnetism: a cautionary note"* y *"Low-Temperature Demagnetization of volcanic rocks containing Multi-Domain magnetic grains: Implications* for the Thellier paleointensity determination", que se incluyen en la **parte II** de este trabajo.

La confiabilidad de muchas determinaciones de paleointensidad puede ser cuestionable, dependiendo de la metodología empleada. Estudios recientes (Goguitachaichvili et al., 1999; Juarez y Tauxe, 2000, Cottrell y Tarduno, 2001, Riisager et al., 2002, Heller et al., 2002, etc.) han mostrado concluyentemente que el mejor método para la obtención de dichas determinaciones es aquel debido a Thellier y Thellier con la inclusión de verificaciones de las magnetizaciones termorremanentes parciales (pTRM check's). Por tal motivo se realizaron dos experimentos de determinación de paleointensidades empleando 1) el método de Shaw (1974) y sus variantes (Kono, 1978; Rolph y Shaw, 1985 y Tsunakawa y Shaw,1994) en muestras de flujos basálticos

Cuaternarios y 2) el método de microondas en muestras históricas del volcán Paricutín, para estimar la confiabilidad y tasa de éxito de estos métodos alternativos. Cabe mencionar en este punto que el primero de los experimentos antes mencionados es una extensión de una investigación previa realizada durante el trabajo de maestría (Morales, J., 1995). Así, se agregaron al trabajo anterior observaciones bajo el microscopio para identificar los minerales opacos portadores de la magnetización, mediciones de susceptibilidad inicial a temperatura ambiente a fin de continuar el monitoreo de posibles alteraciones debidas al calentamiento de los especimenes y se implementó la "prueba de validez" propuesta por Tsunakawa y Shaw,1994. Los resultados preliminares de dichas investigaciones se presentaron en la "Reunión anual de la Unión Geofísica Mexicana" del año 2001 y como resultado de estos estudios surgieron los artículos *"An experimental Re-evaluation of Shaw's Paleointensity Method and its Modifications Using Late Quaternary Basalts"* y *"An attempt to determine the microwave–paleointensity of Paricutín Volcano Lava Flows (Central México)"*, que se presentan también en la **parte II**.

La **parte III** de este manuscrito presenta las aportaciones a la base global de paleointensidades y discute brevemente algunas implicaciones que se infieren de las determinaciones de paleointensidad obtenidas. Se presentan primeramente los resultados de una investigación realizada en rocas volcánicas mexicanas miocénicas, para las cuales se obtuvieron las primeras determinaciones confiables de paleointensidad para Baja California, habiendo obtenido resultados de excelente calidad técnica. Nuevamente, como corolario de este trabajo, los resultados obtenidos y las conclusiones respectivas dieron lugar a la elaboración del artículo *"Geomagnetic field strength During Late Miocene: First Absolute Paleointensity Results from Baja California"*.

Así mismo, a fin de contribuir a disminuir la disparidad existente en el número de determinaciones de PI entre los dos hemisferios [Pick y Tauxe, 1993; Juarez y Tauxe, 2000], la cual impide un análisis fino de los cambios en los parámetros que caracterizan al campo magnético, se prestó particular interés en obtener determinaciones de PI de formaciones localizadas en América del sur, como se propuso en el proyecto de ingreso al doctorado. Para ello se realizaron estudios de PI en la **Formación la Negra** (norte de Chile) y en la **Provincia Magmática de Paraná** (Brasil). Estas dos formaciones son de particular importancia no solo por su distribución geográfica sino también porque permiten cuestionar la validez del así conocido como “bajo dipolar para el Mesozoico”; implicación propuesta primeramente por Prévot et al., 1990 y que asigna una intensidad de campo magnético para este periodo de tan solo una tercera parte del valor actual. Los resultados preliminares de dichas investigaciones se presentaron en la “Reunión anual de la Unión Geofísica Mexicana” del año 2002, mientras que los resultados finales y las conclusiones de estas investigaciones se concretizaron en los artículos respectivos: *“Absolute Paleointensity of the Earth’s Magnetic Field During Jurassic: Case Study of La Negra Formation (Northern Chile)”* y *“On the Reliability of Mesozoic Dipole Low: New Absolute Paleointensity Results from Paraná Flood Basalts”*.

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I. Metodología para la determinación de la paleointensidad absoluta

Antecedentes

La mayoría de los estudios paleomagnéticos realizados hasta la fecha se han centrado en la determinación de paleodirecciones, siendo pocos los dedicados a la determinación de paleointensidades. Lo anterior obedece a que son más los factores que afectan a la intensidad que a la dirección de las rocas magnetizadas, lo que hace la determinación de paleointensidades más difícil y menos confiable que la determinación de paleodirecciones [Coe, 1967]. Se estima que la tasa de determinaciones de paleointensidad fallidas es, en general, del orden del 80% [Kosterov and Prévot, 1998].

La determinación de paleodirecciones se realiza en rocas sedimentarias, ígneas y metamórficas, independientemente de la edad de éstas; sin embargo, la determinación de paleointensidades para períodos mayores a algunos miles de años es usualmente efectuada utilizando rocas volcánicas [Kono, 1978] debido, entre otras cosas, a la gran estabilidad de la magnetización remanente térmica (TRM) ante disturbios externos y a que sus propiedades son relativamente bien conocidas tanto en forma experimental, cuanto en forma teórica [Kono, 1984].

La clave para la determinación de paleointensidades radica en la proporcionalidad existente entre la TRM adquirida y el campo magnético (de baja intensidad) en el cual es enfriado el espécimen [Nagata,1943], ya que la comparación de la magnetización natural remanente (NRM) contra una TRM inducida por un campo magnético conocido proporcionará una estimación de la intensidad del campo antiguo.

En un intento por aumentar la confiabilidad y el número de este tipo de determinaciones, a lo largo del tiempo se han diseñado diferentes métodos para la determinación de paleointensidades. A continuación se describen brevemente.

Método de Könisberger (1938)

En 1938 Könisberg intentó determinar la magnitud del campo magnético antiguo considerando la razón

NRM/TRM

i.e, de la Magnetización Natural Remanente (NRM) a la Magnetización Térmica Remanente (TRM) artificial, adquirida esta última al calentar la roca por arriba de los puntos de Curie de los portadores magnéticos y posteriormente enfriarla en un campo conocido. Lo cual sirvió como base para el desarrollo de métodos más elaborados para la determinación de paleointensidades del campo geomagnético. Esta aproximación posee el inconveniente de que no considera la contribución a la remanencia de las muestras de posibles magnetizaciones secundarias (isotermales, viscosas, etc.).

Método de Thellier & Thellier (1959)

De todos los métodos existentes para la determinación de paleointensidades, aquel debido a Thellier y Thellier es el más empleado. Originalmente éste método fue desarrollado para la determinación de arqueointensidades del campo geomagnético (CGM) en los años 30's.

En este método las muestras son calentadas dos veces a la misma temperatura en presencia de un campo magnético conocido, sin embargo, la posición de las muestras en el segundo calentamiento es opuesta a aquella empleada en el primer paso, de tal manera que las direcciones de los campos aplicados a las muestras son antiparalelas (vistas con respecto a las coordenadas de la muestra). De esta manera, al realizar la semi-suma y semi-resta de los vectores medidos $\mathbf{J_+}$ (= $\mathbf{Y + X}$) y $\mathbf{J_-}$ (= $\mathbf{Y - X}$) (Fig.1.1), se obtienen las coordenadas empleadas en los diagramas de Arai (Fig. 1.2).

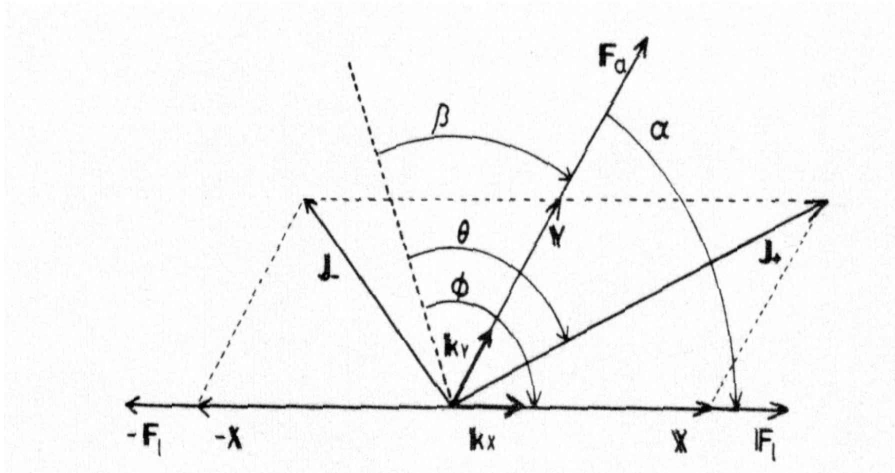


Fig. 1.1. Definición de los diferentes vectores. $\mathbf{K_X}$ es el vector unitario en la dirección del campo de laboratorio $\mathbf{F_l}$ y es por lo tanto paralelo a la componente de la TRM **X**. $\mathbf{K_y}$ es el vector unitario en la dirección del campo antiguo $\mathbf{F_a}$ y es por lo tanto paralelo a la componente de la NRM **Y.** **J+** es la magnetización remanente después del calentamiento a la temperatura T bajo el campo de laboratorio $\mathbf{F_l}$ y es igual a **X+Y.** De forma similar, **J- y Jo** son las remanencias después de calentar a la temperatura T en un campo $\mathbf{-F_l}$ y nulo, respectivamente, e iguales a **Y-X** y **X,** respectivamente (tomada de Kono y Tanaka, 1984).

Esta técnica posee el inconveniente de que un mal alineamiento de los especimenes causado por las imperfecciones de estos puede dar como resultado un aumento en la magnitud de los errores experimentales (Kono y Tanaka, 1984).

Método de Wilson (1961)

Mientras que en la realización del método de Thellier todas las mediciones de la remanencia se efectúan a temperatura ambiente, Wilson emplea una desmagnetización térmica continua, Figura 2a. Para estimar la paleointensidad este autor compara el valor restante de la NRM obtenida de la desmagnetización continua contra aquellos valores obtenidos de la desmagetización térmica continua de la termoremanencia inducida en laboratorio en campos y temperaturas diferentes, Figura 2b.

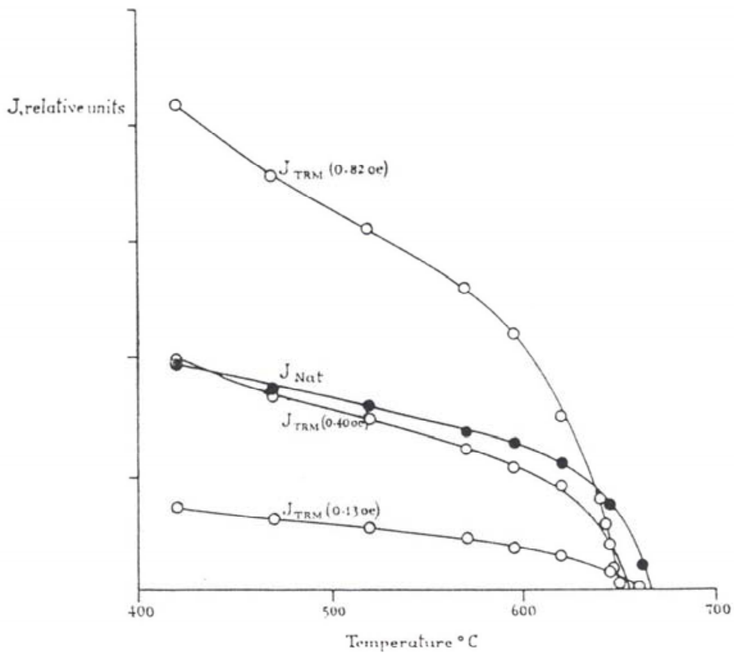


Figura 2a. Curvas de intensidad vs. temperatura de la remanencia natural y de termoremanencias inducidas en el laboratorio con campos magnéticos de 0.13, 0.40 y 0.82 Oe. (Tomada de Wilson, 1960).

Este método posee la gran desventaja de que las muestras deben ser calentadas hasta la temperatura de Curie en campo nulo, para posteriormente efectuar la comparación contra la TRM inducida artificialmente en el laboratoriio. Además de lo anterior, carece de control alguno para detectar posibles alteraciones químicas de las muestras por el calentamiento a que son sometidas.

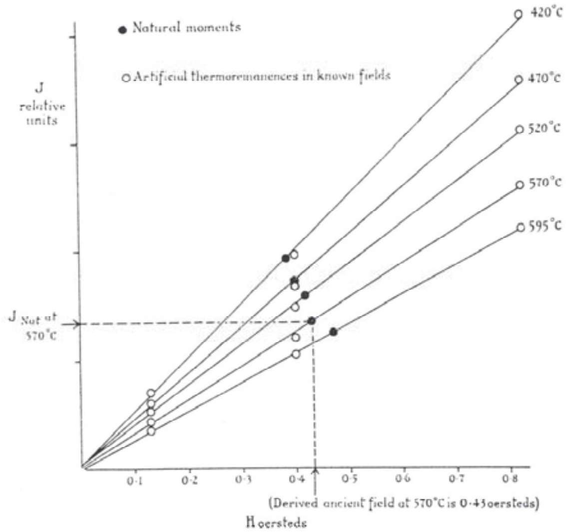


Figura 2b. Los datos de la figura anterior graficados de forma tal, que a una temperatura dada, el momento natural puede ser comparado con las termoremanencias adquiridas en campos conocidos, para así deducir la intensidad del campo antiguo.

Método de Van Zijl (1962)

La premisa fundamental de este método es que para campos de baja intensidad (< 0.8 Oe), la MTR adquirida por una roca enfriada a partir de los puntos de Curie de sus constituyentes magnéticos varia linealmente con la magnitud del campo aplicado [Nagata, 1953; Wilson, 1961]. La utilización del cociente $(NRM/TRM)_{219\ Oe}$, obtenido después de tratar las NRM's y TRM's por campos alternos a un valor pico de 219 Oe, supone que los efectos de las componentes suaves (aquellas con temperaturas de bloqueo o coercitividades bajas) han sido eliminadas [Van Zijl et al., 1962].

Método de Thellier modificado por Coe (1967)

En la variante propuesta por Coe [1967] los especimenes son calentados primeramente en un campo nulo y posteriormente en presencia de un campo artificial de laboratorio, en pasos con temperaturas cada vez mayores. El mal alineamiento de los especímenes deja de ser crítico debido a la simetría axial del campo dentro del horno. Como se mencionó anteriormente, las rocas volcánicas al formarse adquieren una TRM, la cual posee las características siguientes:

i) depende del intervalo de temperatura en el cual ocurre el enfriamiento,

ii) está asociada unívocamente con dicho intervalo y,

iii) es independiente del estado de magnetización fuera del intervalo en cuestión.

Las tres propiedades anteriores dan lugar a la *ley de aditividad de las TRM parciales (TRMP),* enunciada por Thellier en 1938, y que se puede expresar en términos matemáticos como sigue:

$$J(T_n, T_{n-1}) + \cdots + J(T_2, T_1) = J(T_n, T_1)$$

con $T_m \leq T_1 < T_2 < \cdots T_{n-1} < T_n < T_c$

en donde $J(T_i, T_{i-1})$ es la TRMP adquirida al enfriarse la roca desde la temperatura T_i a la temperatura T_{i-1}, mientras que T_m y T_c corresponden a la temperatura del medio ambiente y la temperatura de Curie, respectivamente [Coe, 1967].

Si en una roca la NRM es 100% TRM original, y si el espectro de temperaturas de bloqueo (T_b) de ésta no se altera con el calentamiento, la comparación de la NRM (J_N) contra la TRM (J_A) producida artificialmente en el laboratorio por un campo F_{lab} determina la magnitud del campo antiguo F_{ant} en el lugar y tiempo de formación de la roca, de acuerdo a la expresión

$$J_N(T_1,T_2) / J_A(T_1,T_2) = F_{ant} / F_{lab}$$

donde $J_N(T_1,T_2)$ y J_A (T_1,T_2) son las componentes primarias de la NRM o de la TRM, con temperaturas de bloqueo entre T_1 y T_2. F_{ant} y F_{lab} representan la intensidad del campo magnético antiguo y de laboratorio, respectivamente [Kono, 1978].

Sin embargo, en lugar de evaluar individualmente cada uno de los cocientes dados por las expresiones anteriores, es más común realizar la gráfica de la J_N restante contra la J_A ganada en cada paso de doble calentamiento dando como resultado, en un caso ideal, una línea recta con pendiente negativa como la que se muestra en la figura 1.2.

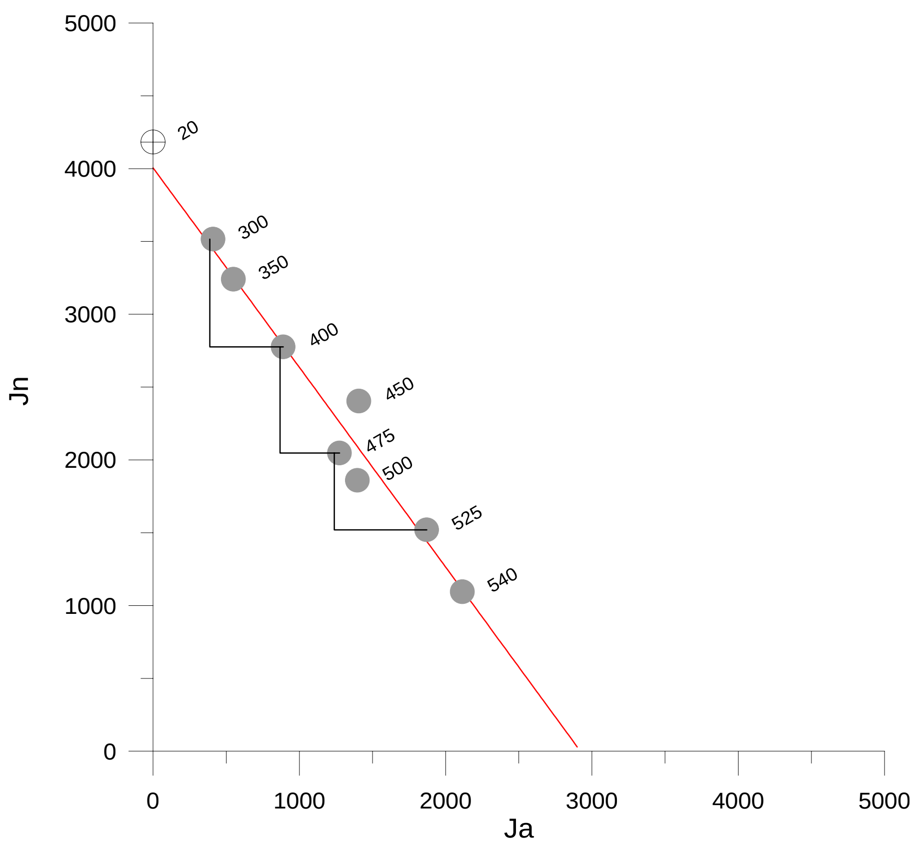


Fig. 1.2. Diagrama de Arai obtenido al realizar el experimento de Thellier y Theller modificado por Coe. Los números a la derecha de los puntos corresponden a la temperatura a la que fueron obtenidos; las rectas en escuadra representan las verificaciones de las TRMP's correspondientes, que se definen más adelante en el texto.

Se dijo "*en un caso ideal",* ya que debido al calentamiento necesario para imprimir la J_A se pueden inducir alteraciones (oxidación-reducción) en los minerales magnéticos originales, lo cual podría producir una disminución o aumento en la capacidad de adquisición de una TRM artificial ó la adquisición de magnetizaciones remanentes secundarias; bajo estas circunstancias, la gráfica es una curva en lugar de una línea recta.

De entre las diferentes variantes del método original de Thellier, el de pasos de doble calentamiento descrito por Coe (1978) permite seleccionar (en la mayoría de los casos) aquel intervalo de temperaturas para el cual el grado de alteración inducido es más bajo.

PROCEDIMIENTO

1.- Se calienta la muestra a una temperatura $T_i > T_m$, se enfría hasta T_m en ausencia de campo magnético y se mide la magnetización restante D_n. Esto elimina la parte de la NRM con temperatura de bloqueo $T_b < T_i$.

2.- La muestra se recalienta a la temperatura T_i y se enfría hasta T_m en presencia de un campo magnético conocido (paralelo al eje del cilindro). Se mide la magnetización resultante D_n + $_p$TRM; La diferencia vectorial entre las dos magnetizaciones anteriores da como resultado la pTRM (J_A).

3.- Los pasos 1 y 2 se repiten para intervalos de temperaturas cada vez mayores hasta alcanzar la temperatura de Curie T_c, eliminando gradualmente componentes más estables de la NRM hasta que su intensidad es pequeña y agregando otras pTRM's.

4. A fin de contar con un medio para detectar alteraciones sufridas por las muestras durante la realización del experimento, se realiza una serie de verificaciones sobre la pTRM’s que consiste en lo siguiente: después de haber efectuado los pasos 1 y 2 a la temperatura T_k se efectúa el paso 1, solo que esta vez a una temperatura menor T_i; posteriormente se realizan los pasos 1 y 2 de la manera descrita arriba a la temperatura T_m. (con T_i <

$T_k < T_m$). En ausencia de alteraciones las pTRM's adquiridas a una misma temperatura T_i deben ser iguales antes y después de haber calentado la muestra a la temperatura mayor T_k.

Lo anterior se representa gráficamente de la forma siguiente: a partir del punto de la gráfica obtenido a la temperatura T_k, se traza horizontalmente una recta con magnitud igual a la de la pTRM generada a la temperatura T_i, y al final de ésta recta se traza una perpendicular cuya magnitud corresponde a aquella de la NRM pérdida a esa temperatura. Si la muestra no ha sufrido alteración apreciable, el final de ésta perpendicular debe caer justo en el punto sobre el cual se esta realizando la verificación (± 15%, valor aceptado ampliamente, e. g. Selkin y Tauxe, 2000); en caso contrario, la muestra se ha alterado y el experimento se termina.

Una vez construida la gráfica J_n - J_A, se ajusta una recta a la serie de puntos obtenidos (Figura. 1.2).

Los errores experimentales aumentan drásticamente si las magnitudes del campo antiguo y de laboratorio difieren mucho entre si (Tanaka y Kono, 1984). Por tal motivo, la selección del campo de laboratorio se debe realizar con base en una estimación de la paleointensidad esperada (si se dispone de ella), o con base en la realización de experimentos piloto en los cuales se emplean un campo de laboratorio de baja intensidad (30 μT), uno de mediana intensidad (60 μT) y otro de alta intensidad (90 μT). En este caso, se elige aquel campo con el cual se obtiene una pendiente cercana a 45°, ya que con esto se disminuyen las incertidumbres asociadas a las coordenadas.

Método de Thellier modificado por Kono y Ueno (1977)

El número grande de calentamientos (y enfriamientos) de una muestra requeridos en los métodos de Thellier y Thellier y Thellier modificado por Coe, antes de que siquiera se pueda decidir si el experimento fue exitoso y que se pueda determinar una paleointensidad, es un serio inconveniente comparado contra el (relativamente sencillo) método de Van Zijl et al. [1962] y otros basados en la comparación de espectros de coercitividades de la NRM y TRM [Kono y Ueno, 1977]. En 1974 Kono [1974] propuso la realización del método de Thellier y Thellier empleando ***un solo*** calentamiento a cada paso de temperatura. El fundamento de esta modificación radica en la suposición de que si la dirección de la magnetización permanece casi constante durante la desmagnetización por campos alternos, ésta permanecerá también constante durante la desmagnetización térmica. Si lo anterior se cumple, los valores de $J_n(T)$ y $J_t(T)$ pueden ser evaluados de una sola medición de la suma vectorial $J_n(T) + J_t(T)$ obtenida en un solo calentamiento, si la dirección del campo de laboratorio F_0 es conocida y no paralela a $J_n(T)$. A fin de reducir los posibles efectos de mal alineamiento de las muestras, y mas aun, para detectar alteraciones mineralógicas de las mismas, se recomienda la aplicación *ortogonal* de F_0. Este método posee el gran inconveniente de que requiere muestras excepcionalmente estables y que los errores experimentales son un tanto grandes [Kono, 1978].

Método de Shaw (1974)

En este método se requiere un solo calentamiento (hasta la temperatura de Curie). En este caso se comparan dos magnetizaciones remanentes anhisteréticas ARM[1], aplicadas antes (ARM_1) y después (ARM_2) del calentamiento requerido, comparación que permite seleccionar la región de fuerza coercitiva H_c en la cual el calentamiento no ha modificado significativamente las propiedades magnéticas de la roca bajo estudio [Shaw, 1974], Nuevamente, en lugar de evaluar individualmente los cocientes correspondientes a cada intervalo de H_c, es práctica común graficar las J_N y la J_A restantes en cada paso de desmagnetización, dando como resultado, también en un caso ideal, una recta con pendiente positiva.

Al igual que en los métodos anteriores, la comparación de una J_N contra una J_A determina la intensidad del campo antiguo, de acuerdo a la expresión:

---

1 ARM: aquella magnetización que se genera al someter a un material ferromagnético a la acción de un campo magnético alterno de amplitud decreciente en el tiempo, simultáneamente a la presencia de un campo magnético constante.

$$J_N(H_1,H_2) / J_A(H_1,H_2) = F_{ant} / F_{lab}$$

donde $J_N(H_1,H_2)$ y $J_A(H_1,H_2)$ son las componentes primarias de la NRM o de la TRM, con coercitividades entre H_1 y H_2. F_{ant} y F_{lab} representan la intensidad del campo magnético antiguo y de laboratorio, respectivamente [Kono, 1978].

PROCEDIMIENTO

1.- Se desmagnetiza la NRM mediante campos magnéticos alternos decrecientes en el tiempo, de valores pico crecientes en cada paso, y se mide la remanencia en cada uno de estos pasos. Los incrementos utilizados varían entre 0.5 y 1 mT, dependiendo del intervalo.

2.- Se genera una ARM_1 utilizando el valor de campo máximo empleado en el inciso 1 y un campo magnético constante de 50 μT. Esta ARM_1 se mide y se desmagnetiza gradualmente de igual forma que en el inciso 1.

3.- Se genera en la muestra una TRM, calentándola por arriba de su temperatura de Curie y enfriándola en presencia de un campo magnético constante de 50 μT. Se mide y se desmagnetiza en forma semejante al inciso 1.

4.- Se genera una ARM_2 utilizando el valor de campo máximo empleado en el inciso 1 y un campo magnético constante de 50 μT. Esta ARM_2 se mide y se desmagnetiza gradualmente de igual forma que en el paso 1.

5.- Con estos datos se construyen las gráficas ARM_1-ARM_2 y NRM-TRM, figura 1.4.

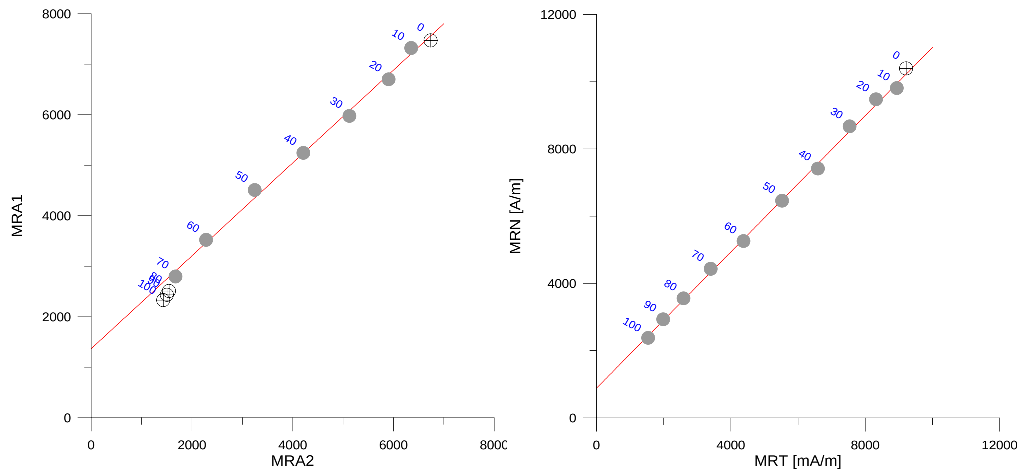


Fig. 1.4. Gráficas de NRM vs TRM y de ARM$_1$ vs ARM$_2$ obtenidas al realizar el experimento de Shaw original.

Una vez construidas las gráficas anteriores, se ajusta una recta a los puntos obtenidos.

Método de Shaw modificado por Kono (1978)

El criterio de aceptación-rechazo tan rígido del método anterior desecha un gran porcentaje de los resultados obtenidos. Posteriormente Kono [1978] presentó una modificación con la cual muchos de los resultados que originalmente fueron desechados pudieron ser considerados como confiables. Dependiendo de la forma de las gráficas ARM$_1$-ARM$_2$ y NRM-TRM éstas se clasifican de la forma siguiente:

1) La gráfica ARM$_1$-ARM$_2$ es lineal y su gradiente es igual a 1

2) La gráfica ARM$_1$- ARM$_2$ es lineal pero su gradiente es diferente de 1

3) La gráfica ARM$_1$-ARM$_2$ no es lineal

y

a) La gráfica NRM-TRM es lineal y pasa por el origen

b) La gráfica NRM-TRM es lineal pero no pasa por el origen

c) La gráfica NRM-TRM no es lineal

La clasificación asignada a una muestra en especial resulta de la combinación del número y la letra apropiada a cada pareja de curvas obtenidas.

La *rigurosidad* atribuida al método de Shaw original se debe a que se descartan todos aquellos experimentos con la excepción de los clasificados como **1a**. Sin embrago, Kono [1978] asumiendo que el espectro de coercitividades de la TRM cambia de la misma manera que aquel de la ARM propuso una corrección de la pendiente de la recta (NRM-TRM) obtenida por el método de Shaw original; a saber,

$$F = F_{lab}*(NRM\text{-}TRM)/ (ARM_1\text{-}ARM_2)$$

La utilización de tal corrección permite que experimentos pertenecientes a las categorías 2a y 2b puedan considerables como exitosos.

Método de Rolph y Shaw (1985)

Rolph y Shaw [1985] retoman la idea anterior y van más allá en un intento por corregir de forma individual los valores de la TRM. La propuesta por consiste en corregir el valor de la TRM por los cambios provocados por el calentamiento de la muestra en la capacidad de adquisición de la TRM en cada paso de desmagnetización. Es decir, en lugar de realizar la gráfica NRM *vs* TRM, ahora se construye la gráfica NRM_i *vs* $(ARM_1/ARM_2)_i$ $*TRM_i$, en donde i es la intensidad del campo de desmagnetización. A diferencia de la corrección propuesta por Kono [1978], esta modificación puede ser usada aun cuando la gráfica NRM-TRM es curva.

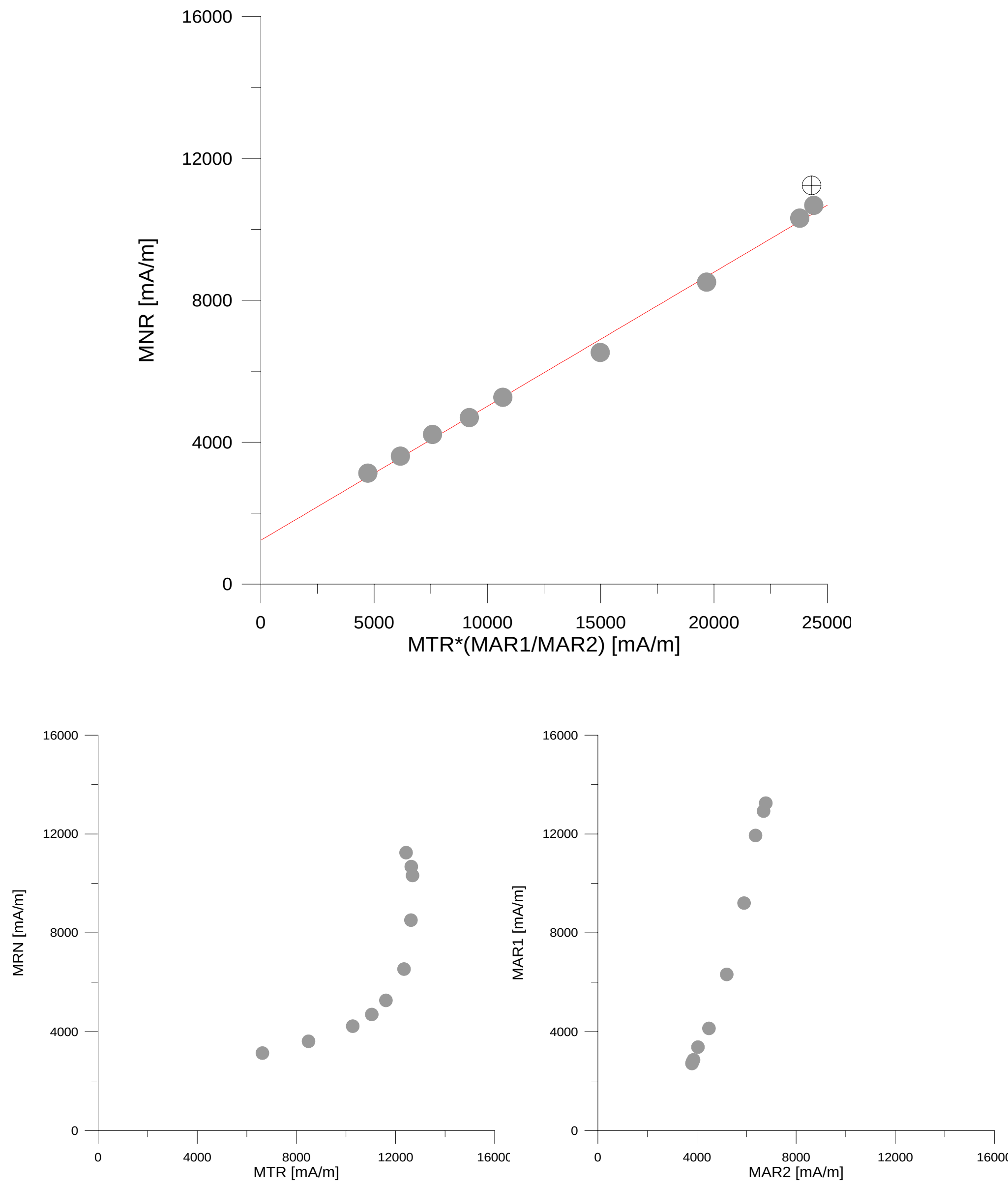


Figura 1.5 Gráficas obtenida al realizar el método de Shaw y Rolph.

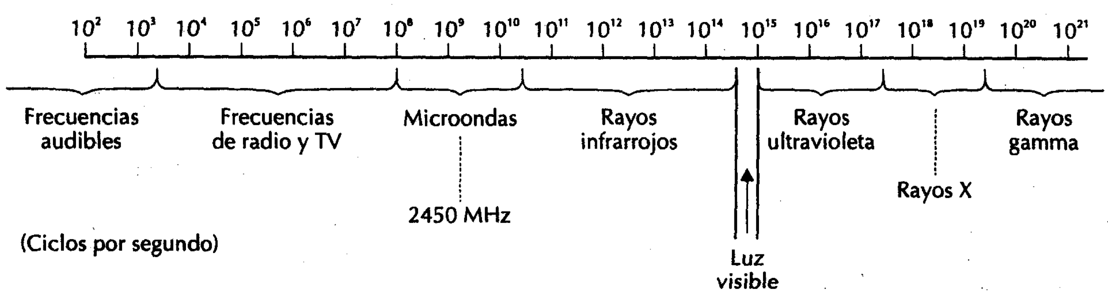
Método de Tsunakawa y Shaw (1994)

Esta modificación es una extensión simple del método de Shaw que utiliza calentamientos dobles de las muestras por arriba de la temperatura de Curie. A partir del último paso del método de Shaw, las muestras son nuevamente calentadas por arriba de la temperatura de Curie y se les induce una nueva TRM (TRM_2) y una tercera ARM (ARM_3) en las mismas condiciones que las primeras. En esta nueva metodología, la TRM_1 se toma como la NRM original a la cual la corrección de Rolh & Shaw (u otra corrección) se le aplica. Si tal corrección es válida entonces se debe obtener un valor correcto para la intensidad del campo de laboratorio. Esto implicaría una gráfica TRM_1 vs TRM_2 corregida (TRM_1-TRM_2*) con pendiente unitaria. Si la diferencia es mayor que el error experimental, se asume que la corrección por ARM no es aplicable y la muestra es por tanto rechazada. El inconveniente de este método es la alteración por temperaturas altas crece de forma *logarítmica* con el tiempo y la alteración debida al segundo calentamiento puede ser mucho menor que aquella durante el primero.

Método de microondas (1993)

La utilización de microondas para desmagnetizar y remagnetizar fue propuesta por Walton et al. (1992, 1993). Posteriormente Shaw et al. (1996) aplican las técnicas anteriores para determinar arqueo-intensidades en cerámicas del Perú. Mediante esta metodología es posible excitar directamente los portadores de la magnetización presentes en una muestra, sin elevar sustancialmente la temperatura de ésta, reduciendo sobre manera la alteración térmica casi inevitable en las técnicas anteriores.

Las microondas son emisiones electromagnéticas de la misma naturaleza que las ondas de radio y televisión, y por lo tanto se ubican en un segmento específico del espectro electromagnético, en el cual se encuentran también la luz visible, la radiación infrarroja, los rayos X, los rayos cósmicos, etc. En la figura siguiente se puede observar la ubicación de las microondas dentro del espectro electromagnético.



De la figura anterior se puede apreciar que, entre las técnicas tradicionales de desmagnetización y remagnetización por calentamiento y la técnica por microondas, existen varios ordenes de magnitud en las frecuencias de las radiaciones empleadas (de 10^8 a 10^{14} Hz). Recordando la relación existente entre energía y frecuencia ($E = h\nu$), es evidente que la radiación infrarroja es mas energética que aquella debida a las microondas; sin embargo, la efectividad para elevar la temperatura de un cuerpo no radica en la intensidad de la energía suministrada si no en la frecuencia de la radiación aplicada, capaz de llevar a las moléculas de un cuerpo a la resonancia. Walton et al. (1991) confirmaron lo anterior al verificar que la temperatura de los granos magnéticos no alcanzaba mas de 300 °C, a pesar de que la potencia suministrada por la fuente de microondas era de 650 W. Por el contrario, 100 W fueron suficientes para alcanzar la temperatura de Curie de los minerales magnéticos cuando la frecuencia de éstas era la adecuada. Así pues, la selección apropiada de la frecuencia de las microondas para calentar exclusivamente las partículas magnéticas en una muestra es de gran importancia.

El calentamiento por microondas es muy rápido y en eso radican algunas de sus ventajas, a saber:

- En experimentos de paleointensidad la alteración térmica de los minerales magnéticos es la mayor fuente de error, y es dependiente del tiempo; de tal forma que calentamientos/enfriamientos rápidos reducirán la alteración térmica.
- La desmagnetización térmica convencional es un proceso destructivo, mientras que la producida por microondas no calienta a las muestras hasta una temperatura suficientemente alta para que esto suceda; por tal razón ésta puede ser aplicada a muestras no adecuadas para la desmagnetización convencional.

En un experimento típico de Thellier las muestras son desmagnetizadas y remagnetizadas a temperaturas progresivamente mayores. En la variante de este experimento empleando microondas la potencia de éstas se incrementa progresivamente por periodos de tiempo fijos. En la figura siguiente se muestra un diagrama de un sistema automatizado para determinación de paleo/arqueo intensidades.

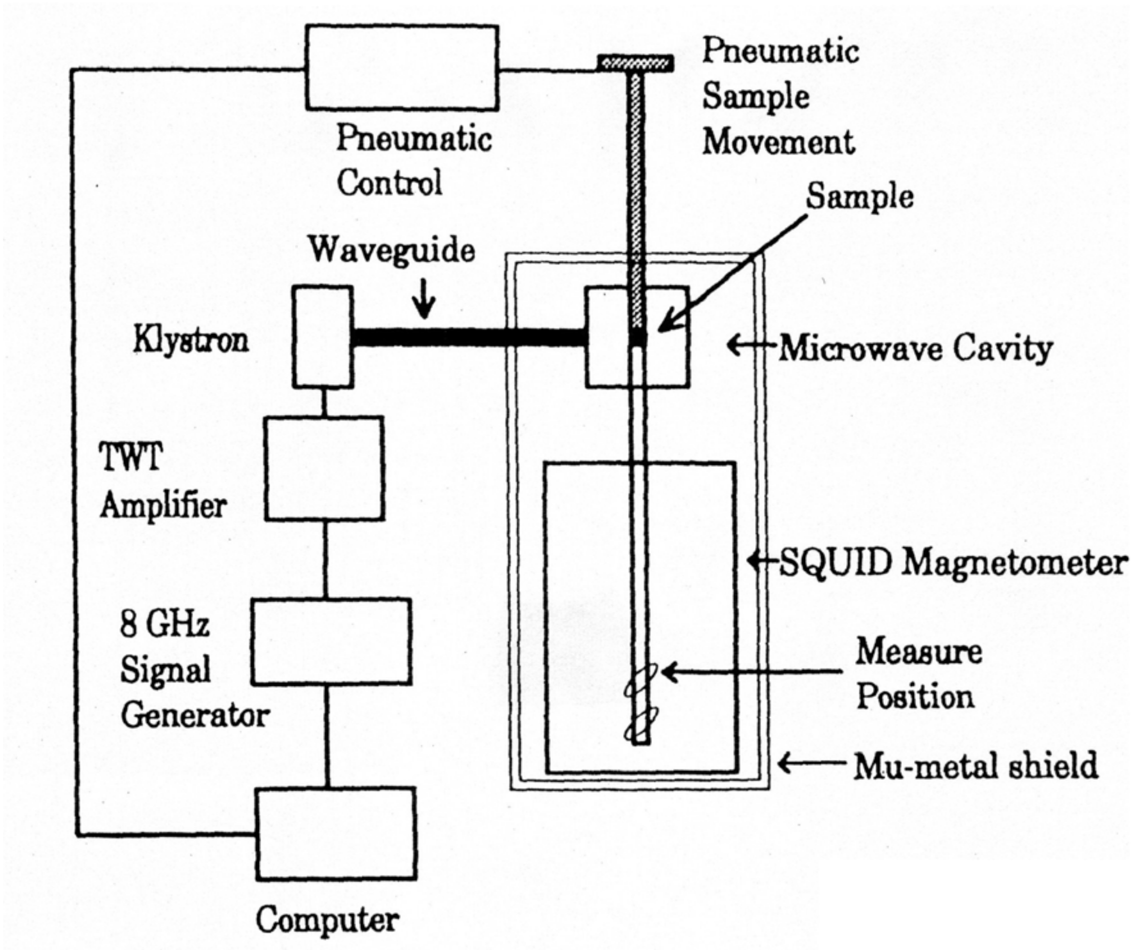


Figura 1.7 Diagrama esquemático de un sistema automatizado de microondas y magnetómetro SQUID. La potencia de salida es controlada por la PC (Tomada de Shaw et al., 1996).

Método de Thellier y Thellier en cristales de plagioclasa (2000)

Se tiene entre la comunidad paleomagnética el consenso, bien justificado, de que el método mas confiable para determinar la intensidad del campo magnético antiguo es aquel debido a Thellier y Thellier [1959] modificado por Coe [1967]. Desafortunadamente, muchas rocas son inapropiadas para la aplicación de ésta metodología debido al crecimiento de nuevos minerales magnéticos asociados, por ejemplo, con la alteración térmica de arcillas en la roca total, Cottrell y Tarduno

[2000]. Los autores anteriores, en busca de una alternativa que evitara las influencias indeseables de la alteración de la roca total, analizaron cristales sencillos de feldespato separados de la matriz arcillosa, los cuales pueden incluir partículas magnéticas capaces de mantener un registro del campo antiguo. Después de ser extraídos los cristales de plagioclasa, se compararon sus propiedades magnéticas contra aquellas de la roca total observándose que las curvas de histéresis magnética eran similares y que sus parámetros se localizaban en la misma región de PSD en un diagrama de Day [1977]. Las curvas termoganéticas para la roca total presentaban cierto grado de irreversibilidad debido a la presencia de titano-maghemita y su subsecuente inversión. Después de seleccionar los cristales con intensidades apropiadas, y de ser colocados en pastillas de sal para garantizar su orientación a lo largo de todo el experimento, se realizaron las determinaciones de paleointensidad siguiendo la metodología propuesta por Coe [1967]. Para el análisis individual de los resultados de las determinaciones obtenidas se eligieron criterios muy rigurosos. El 33% de los cristales de plagioclasa analizados cumplieron con los criterios anteriores. Las determinaciones realizadas en roca total fueron, en general, de baja calidad y menor intensidad. Se atribuye tal diferencia al crecimiento preferencial de nuevos minerales magnéticos en las muestras de roca total; dando como resultado una adquisición de TRM favorecida, hecho reflejado en la poca pendiente en los diagramas de Arai correspondientes.

Discusión

Muchos han sido los métodos y las variantes propuestas a cada uno de éstos, a fin de reducir las alteraciones en la mineralogía magnética causada por el (los) calentamiento (s) a que son sometidas las muestras en los experimentos de paleointensidad. Desde los métodos pioneros de Könisberg [1938] y Van Zijl [1962], los basados en la desmagnetización térmica por pasos de Thellier y Thellier [1959] modificado por Coe [1967], [Kono, 1977]; los que utilizan la desmagnetización por campos alternos de Shaw [1974], [Kono, 1978], [Rolph y Shaw, 1985], hasta la propuesta de Tsunakawa y Shaw [1994], han transcurrido mas de 50 años de investigación. Paralelamente a esto surgen propuestas de análisis de control, criterios de selección de muestras, alertas sobre posibles causas que dan lugar a experimentos fallidos y tratamientos de muestras para aumentar la tasa de éxito de los experimentos de PI y la confiabilidad de los datos obtenidos.

Respecto a esta última cuestión, la estimación de la confiabilidad de los datos de PI se ha basado principalmente en dos tendencias principales: aquella del criterio de *auto-consistencia* (Thomas y Biggin, 2002) de los resultados obtenidos y la basada en *criterios de aceptación* (Selkin y Tauxe, 2000) y magnetismo de rocas. Ninguna de ellas, sin embargo, posee la capacidad de garantizar con cierto grado de confianza que la determinación esté apegada a la realidad, excepto, claro, en los casos de flujos de lava históricos y en el caso de experimentos de laboratorio.

Tomando en cuenta cada nueva determinación de PI no solo incrementa la creciente base de datos globales sino que también contribuye a su dispersión (Pick y Tauxe, 1993), que la gran mayoría de los datos publicados han sido obtenidos mediante métodos carentes de medios para la detección de alteraciones magneto-químicas y que la gráfica de los momentos dipolares virtuales presenta un alto grado de dispersión, valdría la pena, tal vez, reconsiderar los criterios de estimación de confiabilidad puramente estadísticos y trabajar solo con aquellos datos confiables, cualesquiera que sea su cercanía a la realidad. Esto último se discutirá con mayor profundidad en las conclusiones generales.

El método debido a Thellier y Thellier ha probado a lo largo del tiempo ser la metodología más confiable para la determinación de PI y existen varios trabajos que lo validan: demostraciones sobre la validez del método: Kono, 1969, Coe y Grommé, 1973, Tanaka, 1980. Otros tantos que explican las posibles causas de experimentos fallidos y como prevenirlos: Coe, 1967, Levi, 1975, Khodair y Coe, 1975, Kono y Tanaka, 1977, Yongjae y Dunlop, 2002**;** sobre la dependencia con el tamaño de las partículas magnéticas: Levi, 1977 y sobre el rango de aplicabilidad del campo magnético de laboratorio: Tanaka y Kono, 1984.

Aunado a la utilización de criterios de aceptación de resultados de PI, la consideración de factores críticos varios que pueden dar lugar a resultados carentes de significado geomagnético es de gran relevancia. Pick y Tauxe [1993] proponen como un material ideal para determinación de PI a los vidrios basálticos submarinos (SBG), entre otras razones, por su distribución en todo el mundo que ofrece un gran potencial para una mejor distribución de datos de PI en tiempo y espacio, y por ser un material cuyo tiempo de enfriamiento puede ser replicado en el laboratorio.

Kosterov y Prévot [1998], por su parte, llevaron al cabo una serie de experimentos de propiedades magnética y paleointensidades sobre basaltos Jurásicos de la zona de Lesotho, en Sudáfrica, de los cuales es posible obtener paleo-direcciones confiables pero con un comportamiento anómalo al realizar experimentos de Thellier (pendiente muy pronunciada de la curva NRM-TRM a temperaturas intermedias de entre 200 y 460 °C. A partir de los resultados obtenidos de los experimentos realizados, Kosterov y Prévot descartaron como causas de este comportamiento: (1) los efectos de magnetizaciones secundarias y (2) que la magnetización primaria no fuera una verdadera TRM. Asimismo, identificaron tres posibles mecanismos como responsables de este comportamiento:

1. Efectos intrínsecos de multidominio, que no involucran cambios químicos o magnéticos;
2. Algunos cambios químicos irreversibles;
3. Algunos cambios físicos irreversibles

A partir también de los resultados obtenidos, los dos primeros mecanismos fueron eliminados y sugieren que este comportamiento es debido a la transformación de la estructura magnética de los granos de pseudo dominio sencillo (PSD) de una configuración meta-estable a una estable, como resultado de una disminución irreversible en la coercitividad la cual ocurre a temperaturas relativamente bajas (200 - 250 °C).

Por otra parte, Merrill [1970] y Shcherbakova et al. [1996] utilizan la desmagnetización por temperaturas bajas (LTD) como un medio artificial para eliminar la contribución en la remanencia debida a granos de dominio múltiple (MD), concluyendo que dicho tratamiento puede ser usado como limpieza magnética, ya que éste elimina al menos parte de la contribución no ideal de los granos de MD. Así mismo, como un medio para incrementar el éxito de las determinaciones de paleointensidad, Morales et al. [2001] reconocen en esta metodología una herramienta potencialmente útil en la determinación de PI. La aplicación de este tratamiento de temperaturas bajas (desde la temperatura del nitrógeno líquido hasta temperatura ambiente en campo magnético nulo) a magnetizaciones termorremanentes parciales (pTRM's) mostró remover entre un 10 y un 40% de la magnetización portada por granos de dominio múltiple; los cuales no cumplen con las leyes de independencia de Thellier [1959], y son otra causa de fracaso en los experimentos de paleointensidad.

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On the Use of Continuos Thermomagnetic Curves in Paleomagnetism: A Cautionary Note

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Abstract

We report the results of a comparative thermomagnetic investigation on natural volcanic samples. Susceptibility vs temperature (k-T curves), induced magnetization vs temperature and remanent (saturated) magnetization vs temperature continuous curves were recorded on the same virgin samples. Continuous susceptibility curves appear more sensitive to the magnetic mineralogy, in agreement with the theoretical suggestions.

In addition, some new continuous thermomagnetic curves were obtained using VSTM (vibrating sample thermomagnetometer) apparatus, which allows estimating the domain state of magnetic minerals through the study of partial thermoremanent (pTRM) magnetizations. Alternatively, the domain state estimation on the sister samples was derived using hysteresis parameters at room temperature. The interpretation differs depending on method used. This disagreement is probably due to the fact that the domain state estimation using room temperature hysteresis parameters [4] in terms of the plot of magnetization ratio vs coercivity ratio has no resolution for mixture of grain sizes of a single mineral or an assemblage of different minerals. Complex magnetic mineral assemblage with more than one characteristic domain state likely occurs in most natural rocks.

Key Words: Rock-magnetism, Paleomagnetism, Thermomagnetic analyses, Domain structure.

Background

Continuous thermomagnetic curves allow determining the Curie point of magnetic minerals and estimating their thermal stability. Thus, they are indispensable tools in order to identify remanence carriers and to select most suitable samples for absolute paleointensity experiments. Three kinds of thermomagnetic curves are routinely used in paleomagnetism: (1) Induced (saturated) magnetization versus temperature, so-called Js-T curves, (2) Susceptibility versus temperature, k-T curves and (3) Remanent (saturated) magnetization against temperature, Jrs-T curves. Although no Curie temperatures can be determined directly in the later experiment, the trend of unblocking temperatures may be used to estimate this parameter. There is still no definitive agreement among the paleomagnetic community about which method is most appropriate and sensitive, especially in cases when two or more magnetic phases are present. In order to clarify this question, we performed a comparative thermomagnetic investigation recording k-T and Js-T curves on the same volcanic rock samples. In few cases, Jrs-T curves were also obtained using a vibrating sample thermo-magnetometer.

Determination of magnetic domain state is important in paleomagnetism because the stability of magnetic signal depends on the domain structure of magnetic minerals. Preamble detection of multidomain or ‘large’ pseudo-single-domain grains is decisive during absolute paleointensity study since they do not obey Thellier laws of thermoremanent magnetization and no geomagnetic paleointensity could be obtained from this kind of material. Based on the experimental study of the chemically and well-identified synthetic titanomagnetites of known grain size, Day et al. [4] proposed an empiric relation between the domain structure and the hysteresis parameters, which has been widely used in research papers in paleo and rock-magnetism. However, natural rocks almost always plot on the pseudo-single-domain behavior judging from their average hysteresis parameter values [5b].

The magnetic domain structure can be estimated in different ways. Such as blocking and unblocking temperatures of multidomain grains are not equal, the presence of such grains can be detected by means of partial thermoremanence (pTRM) acquisition and demagnetization experiments. A pTRM acquired for example between 300°C and room temperature would not be completely demagnetized below Curie temperature [3, 12].

We applied both the hysteresis and thermomagnetic methods on the same virgin volcanic samples, which contain 'nearly pure magnetite' as showed by previous studies [6, 7, 8]. Thermomagnetic method was found more efficient than regular hysteresis measurements in identifying MD particles.

Magnetic experiments

Low-field susceptibility measurements (k-T curves) under vacuum were performed using a Bartington susceptibility meter (MS-2) equipped with furnace in the paleomagnetic laboratory of University of Montpellier (France). All specimens were heated up to 600°C at a heating rate of 10°C/min and then have been cooled at the same rate. In all cases, the Curie temperatures were determined by the Prévot et al's [11] method.

Induced magnetization in a strong field (B=0.7 T) as a function of temperature, Js-T, was recorded with a Curie balance at the paleomagnetic laboratory of the University of Munster (Germany). These experiments were also carried out under vacuum.

Remanent (saturated) magnetization as a function of temperature, Jrs-T, was recorded using the ‘Orion LTD’ vibrating sample thermomagnetometer (VSTM) at the paleomagnetic laboratory of university of Montpellier. This magnetometer allows to measure the magnetic moment at temperature along one axis. Acquisition of saturation remanent magnetization was realized by a ‘pulse magnetizer’ applying 2.6 Tesla magnetic field.

Hysteresis measurements at room temperature were performed using the AGFM ‘Micromag’ of the paleomagnetic laboratory at Mexico City in fields up to 1T. The saturation remanent magnetization (J_{rs}), the saturation magnetization (J_s) and coercitive force (H_c) were calculated after correction for the paramagnetic contribution. The coercivity of renanence (H_{cr}) was determined by applying progressively increasing backfield after saturation.

The samples used in our experiments come from the volcanic rocks from Southern Caucasus [6], Central Mexico [8] and Philippines [7].

Analyses and Discussion

Both k-T and Js-T curves obtained on the same virgin samples yield identical Curie points and thermal evolutions when magnetic mineralogy is simple i.e. when a unique ferrimagnetic phase is present. This case is represented in Figure 1a, when volcanic samples used in experiment apparently contain ‘almost pure magnetite’. Alternatively, in presence of two or more magnetic phases and thus more complex magnetic mineralogy susceptibility vs temperature curves seem to be more informative (Figure 1b and 1c). In these cases, k-T curves reveal more details of thermal behavior of the samples comparing

Js-T and Jrs-T curves. Moreover, the Curie points determined from different methods are distinct. This slight difference may be owed from the different experimental conditions: k-T experiments take almost 2 hours while Js-T and Jrs-T curves were recorded in about 1 hour. Also, we note that Jrs-T measurements were carried out under air. However, in our case, no direct relation may be established between the experiments under air and under controlled atmosphere. In Figure 1c all three curves are reported on the same diagram. Jrs-T and Js-T curves show only two magnetic phases, when k-T curve yield apparently evidence of three ferrimagnetic components. It is worth noting, that inflection points observed on thermomagnetic curves not always represent Curie points, but may represent some chemical alteration process, this is especially true at low and moderate temperatures. In any case, k-T curves seem to be more sensitive for thermal evolution of natural samples.

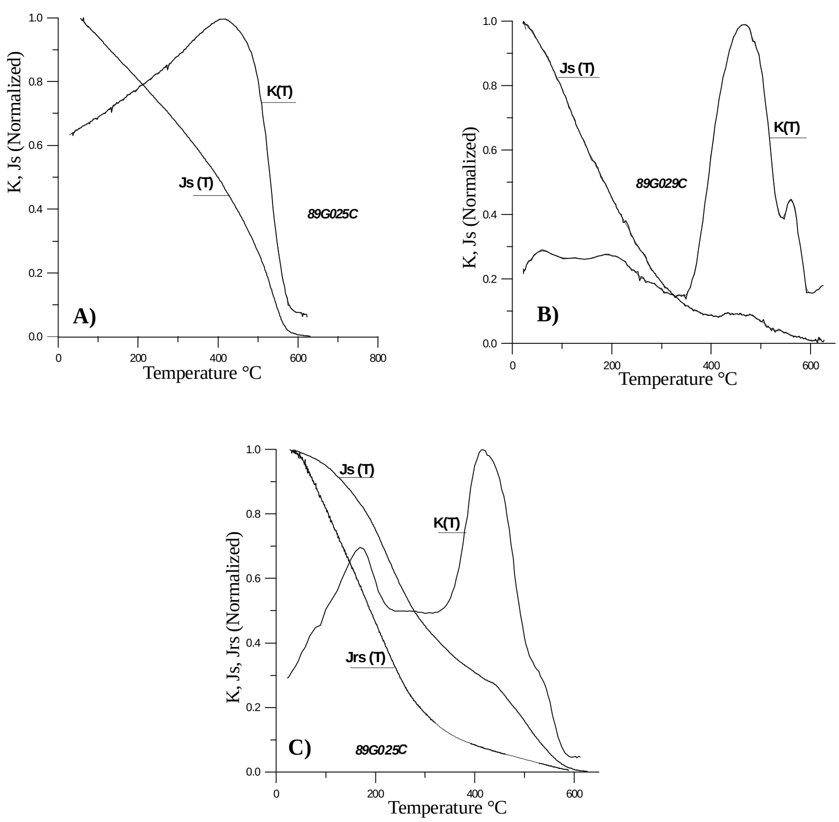
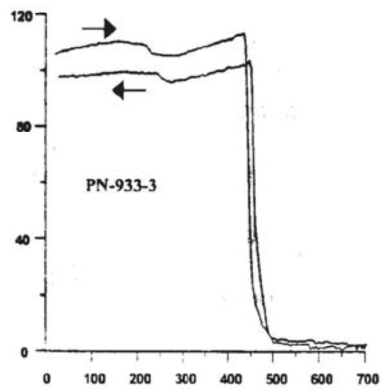


Figure 1. Examples of susceptibility (k-T), induced magnetization (Js-T) and remanent (saturated) magnetization (Jrs-T) vs temperature curves performed on the same virgin volcanic samples (see also text).

This is in agreement with theoretical calculations. Susceptibility depends on both spontaneous magnetization (Ms) and coercive force (Hc) [*for single domain grains k=0.349(**Ms/Hc**) after Dunlop* [5]*; for multidomain magnetic grains k depends on the demagnetizing factor N* [9, 10, 13], *k ~ 1/N, when N varies from 3.8 to 3.9 following Banerjee,* [2]. *An empiric relation was established between susceptibility and temperature: k = (**Ms/H**)xL(v**MsH**/kT) where v is a volume of magnetic grain, k is Boltsman constant, T is temperature, L(f) is Langevin function L(f) = f if f<0.2 and L(f)=1 if f>0.2*]. Thus this relation combines the thermal variation of two magnetic parameters, when induced magnetization (when saturation is reached) describes the thermal evolution of spontaneous magnetization only.

Further evidence of 'superiority' of k-T curves is shown on Figure 2. The pumice (dacite) sample comes from Pinatubo 1991 eruption and contains two magnetic phases as previously identified by Bina et al. [1] and Goguitchaichvili and Prévot [7]: the ilmeno-hematite with y=0.53 and titanomagnetite with x= 0.1. Susceptibility vs temperature curve revealed the presence of both ferrimagnetic minerals (Curie point around 250°C for ilmeno-hematite) when it is completely masked in the Js-T curve. This later only yields evidence for Ti-poor titanomagnetites. This is probably because in high fields the magnetic signal of ilmeno-hematites is negligible with respect of titanomagnetites. Thus, precise determination of magnetic mineralogy requires k-T curves.



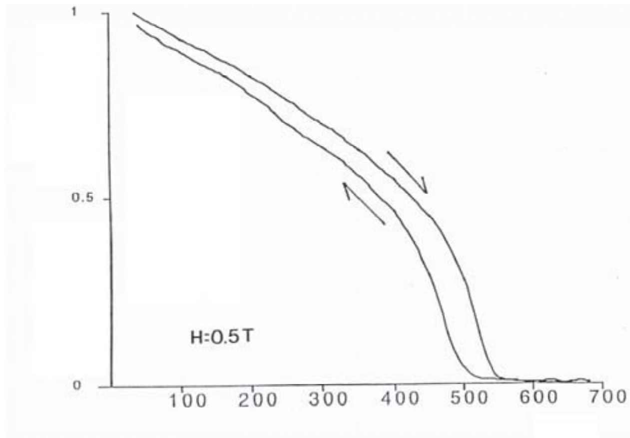


Figure 2. Susceptibility (left) and induced magnetization vs temperature curves from the volcanic (dacite) sample from Pinatubo. Redrawn from by Bina et al., 1999.

Vibrating sample thermomagnetometer is a useful tool to investigate the properties of partial thermoremanent magnetization, as it allows the acquisition and demagnetization of TRM or pTRM, and the measuring of remanence during thermal demagnetization in a continuos way along one axis. For our experiments, the magnetization was measured and fields were applied along the axis of maximum magnetization. We selected 11 representative samples, which show reversible behavior during k-T measurements with a Curie point, near that of pure magnetite. First their natural remanent magnetization (NRM) was demagnetized,by heating the samples up to 600°C, and then were cooled down to 300°C in absence of magnetic field. At this temperature a magnetic field of 50 microTesla was switched on and maintained during the cooling cycle up to room temperature. Thus, a pTRM was created between 300 and 25°C (pTRM up after Shcherbakova et al. [12]), which was subsequently thermally demagnetized (Figure 3, left side). The amount of remanent magnetization still present after heating above the highest pTRM acquisition (300°C in our case) can provide information about the fraction of multidomain grains in a sample [6, 12].

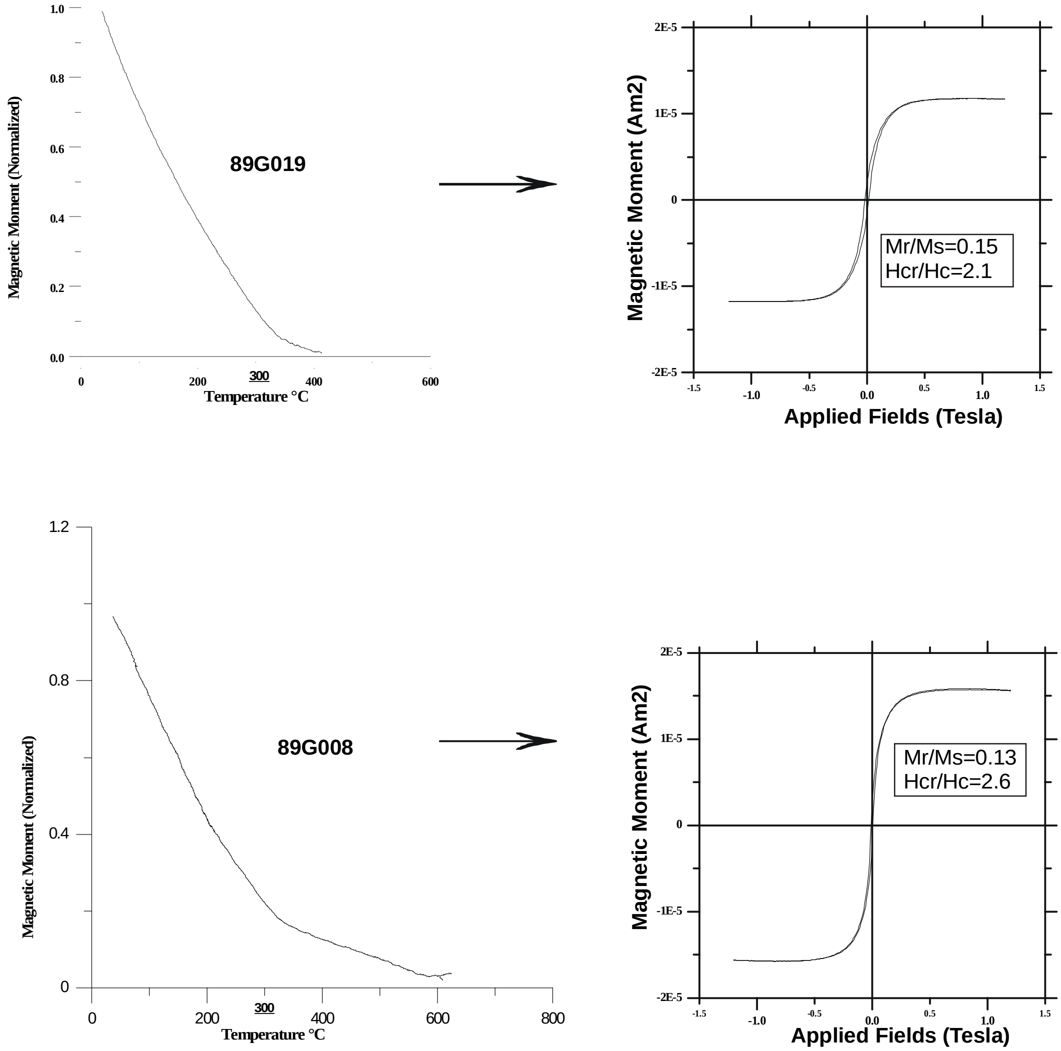


Figure 3. Continuos thermal demagnetization of a partial thermoremanent magnetization acquired from 300°C to 28°C in a magnetic field of 50 μT and corresponding room temperature hysteresis cycle (corrected for paramagnetism).

In five of eleven curves, pTRM blocking temperatures were quite similar to their blocking temperatures (Figure 3, upper part). This result probably indicates the presence of dominantly single-domain magnetic grains. In six other cases, samples had to be heated to 600°C to completely demagnetize the pTRM remainded undestroyed (Figure 3, lower part). In those cases, a significant fraction of grains seems to have a multidomain structure. Hysteresis measurements in both cases indicate to pseudo-single-domain grain region judging from the values of Ms/Mrs and Hcr/Hc [4]. It should be noted that this is commonly the case when natural rocks are analyzed. On other hand, the samples used in hysteresis experiments are relatively small and may be not representative of the whole rock. In our case, however, the difference between hysteresis parameters ($J_{rs,}$ J_s, $H_{c,}$ H_{cr}) of small samples belonging to the same core was insignificant.

Several problems arise when using the Day diagram: 1) Natural rocks are complex magnetic systems, which contain grains of variable size, coercivity and composition. Thus, generalizations based on studies of synthetic/crashed materials with established chemical composition are probably of limited use. 2) Natural rocks may contain solid solutions like ilmeno-hematites, titanohematites or titanomaghemites. In these cases the Day plot is of limited value. And 3) When superparamagnetic grains are present, they can contribute in induced magnetization deviating artificially hysteresis parameters towards pseudo-single or multidomain region on the Day diagram. Keeping this in mind, the domain structure estimation through partial thermoremanence is probably more precise, although requires much more laboratory time.

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Low-Temperature Demagnetization of Volcanic Rocks Containing Multi-Domain Magnetic Grains: Implications for the Thellier Paleointensity Determination

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ABSTRACT

Thirteen samples of volcanic rocks from Caucasus, Iceland and Philippines were subjected to low-temperature treatment in order to remove the part of remanent magnetization carried by multi-domain grains. Ten to 40% of multi-domain partial thermoremanence were removed, which may help to increase the success of paleointensity experiments. A sample from the 1991 eruption of Pinatubo volcano shows a total self-reversal due to ilmeno-hematites and partial self-reversal between 500° and 575 °C due to titano-magnetites. Low temperature demagnetization applied to self-reversed remanence suggest that the high-temperature peak is almost suppressed by this treatment.

KEY WORDS: Rock-magnetism, Low-temperature demagnetization, Domain structure.

RESUMEN

Seleccionamos trece muestras de unidades volcánicas previamente estudiadas de los Cáucasos, Islandia y las Filipinas para usarse en este estudio de estructura de dominio magnético y desmagnetización de baja temperatura. Se aplicó tratamiento de baja temperatura a las magnetizaciones termoremanentes parciales para remover la parte de magnetización remanente portada por granos multi-dominio. En general, 10 a 40% de la ‘termoremanencia parcial multi-dominio’ fue removida con este tratamiento, esto puede ayudar a incrementar el éxito de las mediciones de paleointensidad. Las muestras de Filipinas provienen de la erupción del volcán Pinatubo y muestran un fenómeno de auto-inversión total debido a la presencia de ilmeno-hematitas y una auto-inversión parcial entre

500° y 575°C debido a las titanomagnetitas. La desmagnetización de baja temperatura aplicada por primera vez a remanencia auto-inversa muestra que el pico de alta temperatura es casi eliminado por el tratamiento.

PALABRAS CLAVES: Magnetismo de rocas, Desmagnetización de baja temperatura, Estructura de dominio.

INTRODUCTION

Reliable absolute paleointensity results are generally much more difficult to obtain than reliable directional data. Only volcanic rocks that satisfy certain specific magnetic criteria can be used for paleointensity determination (Kosterov and Prevot, 1998). Thellier and Thellier (1959) provided a method to determine the intensity of the geomagnetic field on volcanic rocks and archeological materials carrying thermoremanent magnetization (TRM). This method is considered to be the most reliable. However, several conditions have to be obeyed to ensure the significance of the paleointensity results:

1. The primary remanent magnetization must be a TRM and must not have decayed significantly.
2. Secondary components must be weak with respect to the primary component and must be removed at relatively low temperatures.
3. The remanence must be carried mainly by non-interacting single-domain magnetic grains to ensure the independence of the partial thermoremanent magnetizations (pTRM) (Thellier and Thellier, 1944, 1959).
4. No chemical/magnetic changes occur during laboratory heatings.

Conditions 1 and 2 are fulfilled by a significant fraction of volcanic rocks. The magnetic carriers in volcanic rocks selected for paleointensity experiments is commonly a Ti-poor titano-magnetite spinel formed from spinodal decomposition of an original Tirich titano-magnetite. The grain size of this spinel phase is generally larger than single-domain/pseudo-single-domain (SD/PSD) threshold (Kosterov and Prévot, 1998). This makes impossible the determination of geomagnetic paleointensity for most volcanic rocks.

The present work is an effort to try to eliminate the fraction of the remanence carried by multi-domain magnetic grains and thus, to increase the success of Thellier paleointensity determination by low-temperature demagnetization.

THEORETICAL BACKGROUND

Low-temperature demagnetization (LTD) is the process of cooling a sample to the isotropic temperature T_I = 120-135 K of magnetite (Bickford et al., 1957; Syono, 1965). Usually the sample is cooled to liquid nitrogen temperature (77 K) and then warmed to room temperature in zero field. This is an effective means of erasing lower coercivity remanence (Nagata, 1961; Ozima et al., 1964; Merrill, 1970). However, LTD cannot be carried out in stepwise fashion to yield a record of the progressive removal of soft component magnetization.

In theory, LTD should destroy the remanence carried by multi-domain grains of magnetite or Ti-poor titano-magnetite (Markov et al., 1983). Around 130 K, the magnetocrystalline anisotropy K_1 changes its sign and the easy magnetization axis changes orientation (Bickford et al., 1957). This transition is accompanied by abrupt changes in coercivity, remanence and susceptibility (Aragon, 1985, 1992). The magnetic memory (the fraction of TRM surviving after LTD) is much more resistant to AF demagnetization than original TRM before LTD (Heider et al. 1992). This hypothesis is supported by McClelland

and Shcherbakov (1995). They demonstrated that LTD could destroy the component of multi-domain remanence. In this work, we examined whether technique involving low-temperature demagnetization could be used to remove the remanence carried by multi-domain grains in natural volcanic rocks.

EXPERIMENTAL RESULTS

Remanence measurements of different rock samples were recorded on ‘Orion Ltd’ vibrating sample thermomagnetometer (so-called VSTM) at the paleomagnetic laboratory of the University of Montpellier. This magnetometer allows the acquisition and demagnetization of TRM and pTRM (partial thermoremanent magnetization), and the measuring of magnetic moment during thermal demagnetization in a continuos way along one axis. The sensitivity of the magnetometer is $5*10^{-10}$ Am^2, the maximum field H which can be applied is $4*10^3$ Am^{-1} and residual field after turning off the magnet is < 0.1 Am^{-1}. For our experiments, small cylindrical specimens were cut. Their magnetization was measured and magnetizing fields were applied along their axis of maximum magnetization.

We selected twelve representative volcanic (basalts) sample coming from Southern Caucasus and Iceland. They yielded a stable remanence and reversible behavior during continous susceptibility vs temperature measurements with Curie points near pure magnetite as showed by previous studies (Goguitchaichvili et al., 1997, 1999a and b, 2000b). One more sample belongs to Pinatubo (Philippines) 1991 eruption. It contains two magnetic phases: ilmeno-hematite with y=0.53 and titano-magnetite with x=0.09 (Goguitchaichvili and Prévot, 2000a).

Determination of the viscosity index (Thellier and Thellier, 1944; Prévot, 1983) is useful to obtain information about paleomagnetic stability of the samples. We placed all samples during 15 days with one axis aligned with Earth's magnetic field. After measuring magnetization (**M_d**), samples were placed for 15 days in a field-free space, and the magnetization (**M_0**) was measured again. The viscosity index is $V = [(Z_d - Z_0) : M_{nrm}] \times 100$, where Z_d and Z_0 are the magnetization components of M_d and M_0 which are parallel to the magnetizing field and **M_{nrm}** is the intensity of natural remanent magnetization. The samples did not present a large capacity for viscous remanence acquisition. Viscosity index was generally less than 5%, a value which is low enough to obtain precise measurements of the remanence during the process of thermal demagnetization (Prévot et al., 1985).

The samples belong to sites with very low angular standard error. All values of α_{95} were within 5° of cleaned natural remanent magnetization (NRM). They carry essentially a stable, single-component magnetization, observed upon both thermal and alternating field (AF) treatments (Figure 1). A minor secondary component, probably of viscous origin, is easily removed by low temperatures / AF fields. The median destructive fields range mostly in the 20-25 mT interval, suggesting the existence of pseudo-single domain (PSD) grains as remanence carriers (Dunlop and Ozdemir, 1997). This behavior might also be due to a mixture of single-domain (SD) and multi-domain (MD) magnetic particles. The major part of remanence is destroyed at 500-550°C, suggesting low-Ti titanomagnetite as responsible for magnetization.

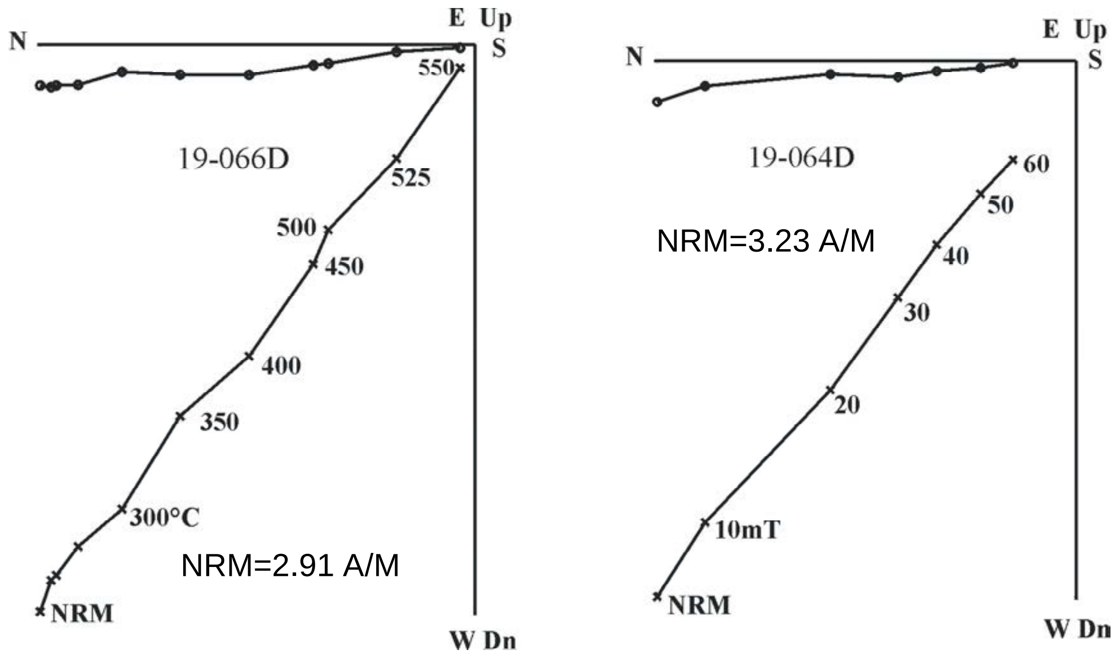

Figure 1. Orthogonal vector plots of stepwise thermal or alternating field demagnetisation of selected samples. The numbers refer either to temperatures in °C or to peak alternating fields in mT. o – projections into the horizontal plane, x – projections into the vertical plane.

Continuous low-field susceptibility measurements were performed in vacuum using a Bartington MS2 susceptibility meter with furnace. The selected samples show a single ferrimagnetic phase with a Curie point compatible with Ti-poor titanomagnetite (Figure 2a). Some of the curves are not perfectly reversible.

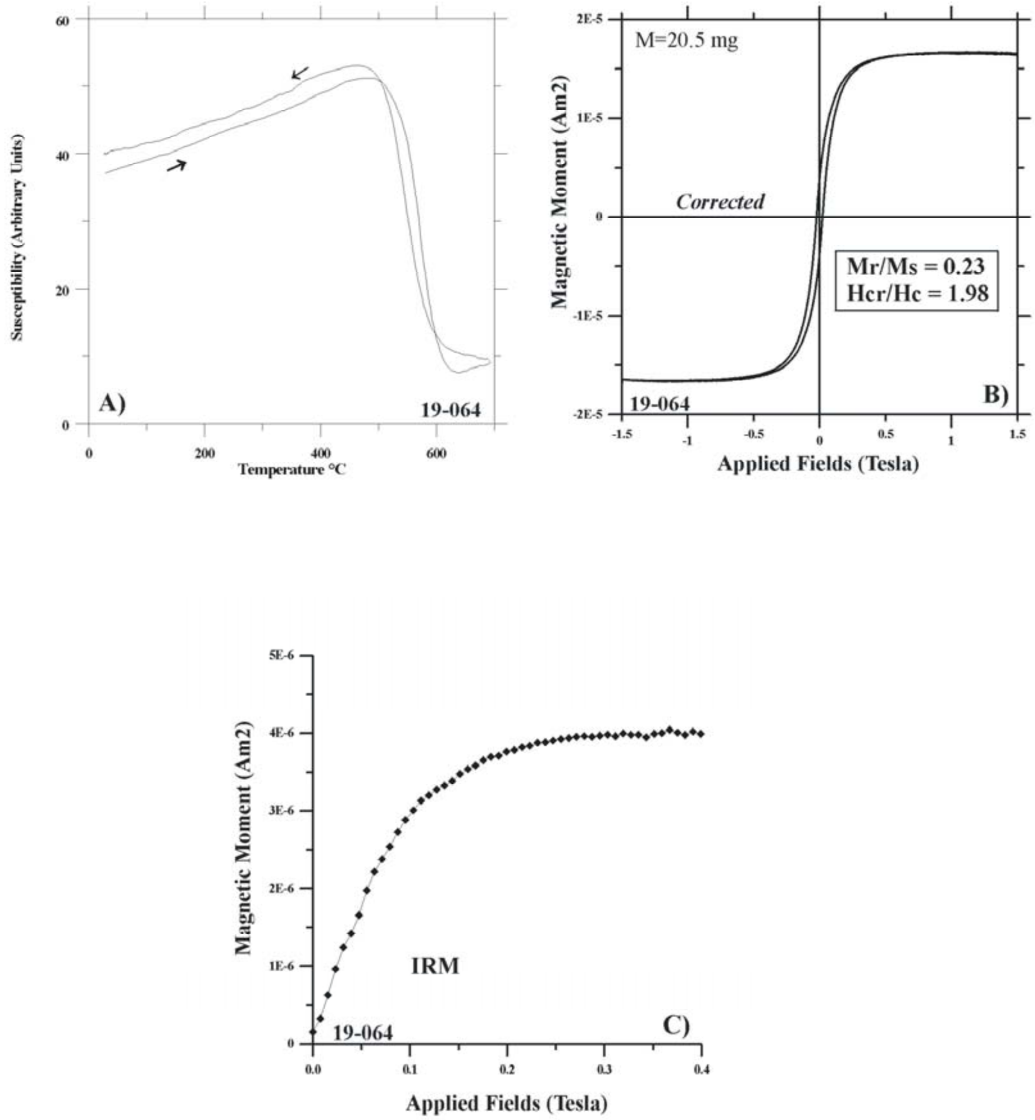

Figure 2. a) Representative susceptibility versus temperature curves.The arrowsindicate the heating and cooling curves (see also text). b) Typical examples of hysteresis loop (corrected for dia/paramagnetism) of small chip samples and c) isothermal remanence (IRM) acquisition curve.

Hysteresis measurements at room temperature using an AGFM-Micromag apparatus were carried out in fields of up to 1.4 Tesla. The hysteresis parameters (saturation remanent magnetization Jrs, saturation magnetization Js, and coercive force Hc) were calculated after correction for the paramagnetic contribution. Coercivity of remanence Hcr was determined by applying a progressively increasing back field after saturation. IRM (isothermal remanent magnetization) curves show that saturation is reached in moderate fields of the order of 150-200 mT (Figure 2C), which points to some spinels as remanence carriers. Judging from the ratios of hysteresis parameters (Figure 2B), it appears that the samples fall in the PSD grain size. This probably indicates a mixture of MD and SD grains. If some superparamagnetic grains are also present, the measured coercive force and saturation magnetization may be underestimated.

As blocking and unblocking temperatures of multidomain grains are not the same, the presence of such grains can be detected by means of pTRM acquisition and demagnetization experiments. A pTRM acquired between 300°C and room temperature, would not be completely demagnetized below the Curie temperature (Bolshakov and Shcherbakova, 1979; Worm et al. 1988). This method is more efficient than regular hysteresis measurements to detect multi-domain grains, but requires much more laboratory time (Goguitchaichvili et al. 2001).

The experimental procedure was as follows:

First, the NRM of the selected samples was demagnetized using VSTM (Figure 3, curve 1). Next they were cooled to room temperature in a 50 μT field, so that a TRM was acquired. Afterwards, this TRM was demagnetized (Figure 3, curve 2). Subsequently, pTRM was given to each of the samples between 300 and 25°C using the same laboratory field intensity, which was subsequently thermally demagnetized (Figure 3, curve 3). The amount of remanent magnetization still present after heating above the highest pTRM acquisition temperature (300°C in this case) can provide information about the fraction of multidomain grains in a sample (Shcherbakova et al., 1996). In a final step, the same pTRM was given again to those samples, where a significant multi-domain-grain fraction seemed to be present, and they were then placed into liquid nitrogen and left in free magnetic space approximately for two hours. Subsequently, they were thermally demagnetized (Figure 4, curve 4).

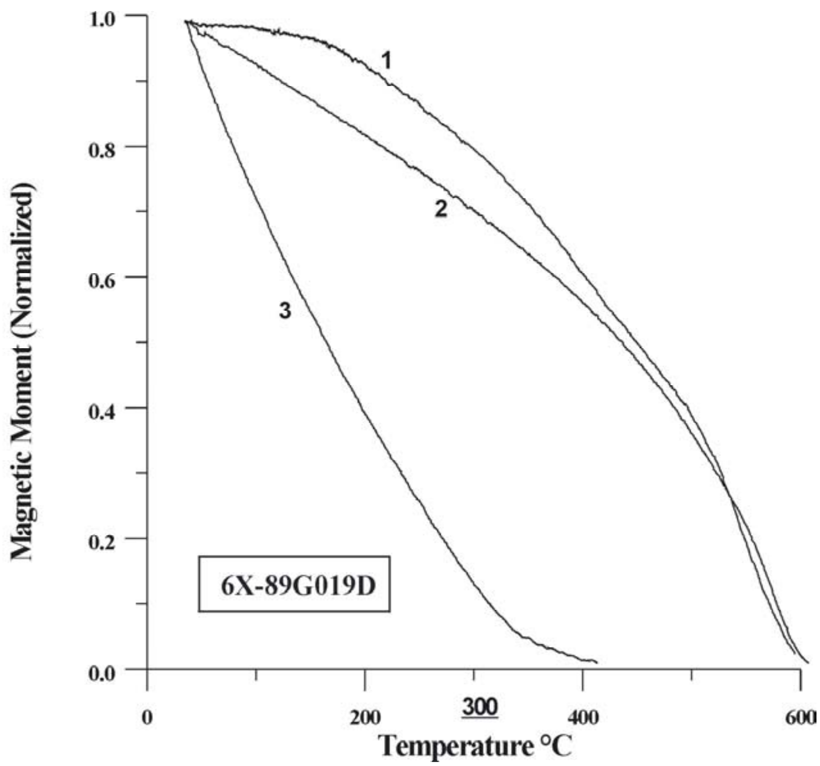

Figure 3. Continuos thermal demagnetization of natural remanent magnetization (curve 1), thermoremanence (curve 2) acquired in a field of 50 μT, and partial thermoremanence (curve 3) acquired in a field of 50 μT, cooling down from 300°C to room temperature (Retrieved from Goguitchaichvili et al. 2000b).

Discussion

In five of twelve curves, pTRM laboratory unblocking temperatures were quite similar to their blocking temperatures. Thus the amount of pTRMs remaining undestroyed above 300°C was less than 15% of the total pTRM (Figure 3). This result probably indicates predominantly single-domain magnetic grains (Goguitchaichvili et al., 2001). The seven other samples had to be heated to 600°C to completely demagnetize the pTRM (Figure 4). In those cases, a significant fraction of grains may have a multi-domain structure.

Low-temperature treatment removed a multidomain pTRM only partly (Figure 4, Table 1). We observed the reduction of pTRM tail from about 10 to 40%. This may be due to the fact that small deviations from stoichiometry of magnetite have an important effect on the Verwey transition, and the magnetic memory is controlled in part by the internal stresses developed during partial oxidation (Ozdemir et al. 1993).

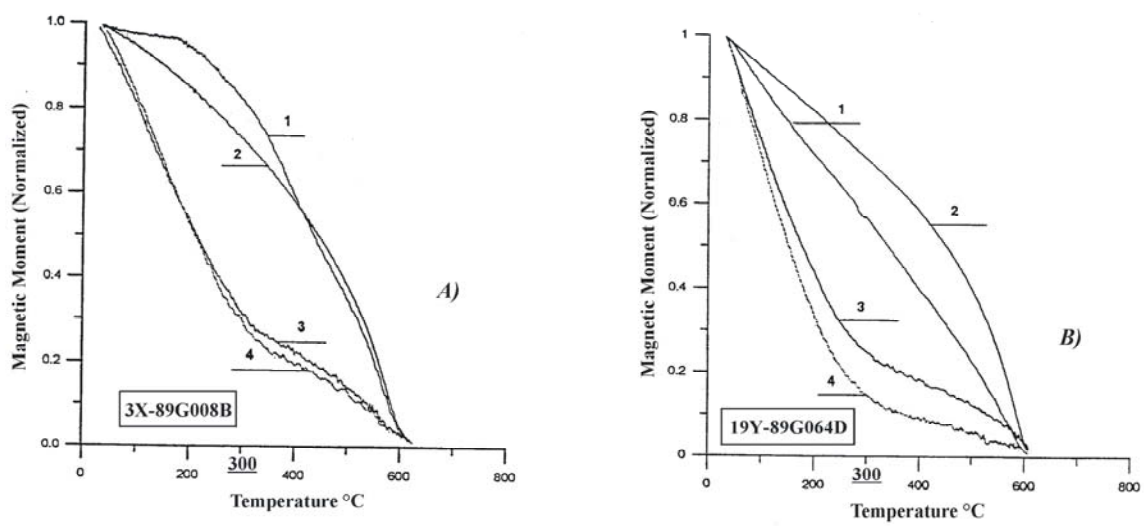

Figure 4. Notation as in Figure 3. Curve 4 is a continuos thermal demagnetization of partial thermoremanence acquired in a field of 50 μT, from 300°C to room temperature after treatment with liquid nitrogen (See also text).

Figure 5 show the thermal demagnetization of a pTRM (600°C, 50 μT, 550°C), acquired by cooling down from 600 to 550 °C in the presence of a laboratory field of 50 μT, carried by a sample containing: intermediary ilmeno-hematite and Ti-poor titano-magnetite before (curve 1) and after LTD treatment (curve 2). Both minerals have a multidomain domain structure as showed by Bitter technique observation (Bina et al., 1999). The sample came from the 1991 eruption of Pinatubo volcano and showed total self-reversal due to the presence of ilmeno-hematites, and partial self-reversals observed from 500°C to 575°C due to titano-magnetite (Figure 5). This may be the first time that LTD was applied to self-reversed thermoremanence. In this case, the high-temperature peak was almost entirely suppressed by LTD.

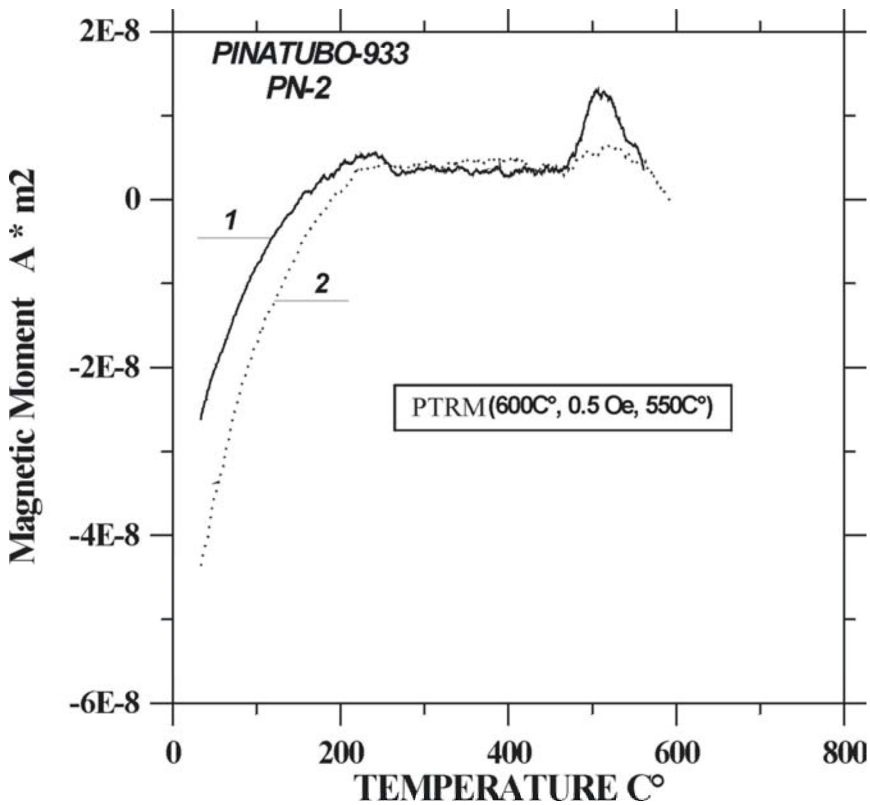

Figure 5. Continuos thermal demagnetization of partial thermoremanence acquired in a field of 50 μT, cooling down from 600°C to 550°C before (curve 1) and after (curve 2) LTD treatment (see also text).

Sample	pTRM tail	pTRM$_1$(300°C)	pTRM$_2$ (300°C)	pTRM$_2$/pTRM$_1$
	(%)	**(A/m)**	**(A/m)**	
3X-89G008B	31	0.34	0.31	**0.91**
19Y-89G064D	26	0.46	0.29	0.62
FL2_3	19	0.14	0.12	**0.86**
FL1_1	24	0.78	0.55	**0.70**
FL3_7	24	1.51	1.13	**0.75**
KY3_7	21	1.05	0.78	**0.74**
SE1_1	28	0.67	0.54	**0.81**

Table 1. The magnetic parameters of partial thermoremenent magnetization (pTRM) of samples which yielded 'multi-domain like' behaviour (see also text). pTRM$_1$ – intensity of partial thermoremanent magnetization formed from 300°C to room temperature in a field of 50 μT. pTRM$_2$ – *idem*, after LTD treatment (see also text). pTRM tail – the amount (in %) of pTRM (300°C, 50μT, T_0) remainig undestroyed above 300°C.

Conclusion

Thirteen samples selected from previously studied volcanic units from Caucasus, Iceland and Philippines are tested. Samples present small viscosity indexes, Curie points compatible with Ti-poor titanomagnetites, stable remanence and reversible behaviour of susceptibility-temperature curves, and hysteresis parameters compatible with the pseudo-single-domain range (probably mixtures of single-domain and multi-domain states).

The remanence carried by multidomain grains is removed only partly by low-temperature demagnetization. In general, 10 to 40% of 'multidomain pTRM remanence' were destroyed. This result may help to increase the success of paleointensity measurements.

Acknowledgment

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An Experimental Re-evaluation of Shaw's Paleointensity Method and its Modifications Using Late Quaternary Basalts

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Abstract

Absolute paleointensity experiments were carried out using Shaw's method (1974) and its modifications (Kono, 1978; Rolph and Shaw, 1985 and Tsunakawa and Shaw, 1994) on 49 samples belonging to six Late Quaternary basaltic flows from central Mexico. Samples were selected from a large collection because of their low viscosity index, stable remanent magnetization and close to reversible continuous thermomagnetic curves. Moreover, they previously yielded high technical quality Thellier paleointensity results, which makes them good candidates to asses the reliability of Shaw's paleointensity method. Only 13 samples yielded acceptable results using Shaw's original method (ARM_2/ARM_1 ratio varies from 0.95 to 1.05 for accepted determinations); though 10 samples do not pass the validity test proposed by Tsunakawa and Shaw (1994) and thus should be rejected for paleointensity analyses. Rolph and Shaw's (1985) method gives reliable determination only in one case and no single determination was obtained by Kono's (1978) modification. Our results indicate an extremely low success rate of Shaw's paleointensity method, which may be due to magneto-chemical changes occurred during heating of samples above its Curie temperature.

Introduction

A better understanding of the physics of the Earth's deep interior and the functioning of the geodynamo can be achieved by the study of absolute paleointensity variation (Coe et al., 2000). However, the number of this type of determinations is still low, and the reliability of some of them is questionable, because of the methodology employed (e.g., Goguitachaichvili et al., 1999; Riisager et al., 2002). The method developed by Thellier and Thellier (1959), based on the characteristics of thermoremanent magnetization (TRM) acquired by cooling on a weak field in nature and the laboratory, remains a reliable and widely used method for paleointensity determination. Unfortunately, change in TRM acquisition capacity resulting from alteration due to various laboratory-heating steps to progressively higher temperatures results in small success rate in the Thellier method. This has encouraged development of alternative methods requiring a small number of heating steps or using other remanent magnetizations. Methods that have received wide attention are those based on the similarities in coercivity spectra of anhysteretic remanent magnetization (ARM) as compared to TRM (Shaw, 1974; Dunlop and West, 1969).

The Shaw paleointensity method (1974) uses alternating field (AF) demagnetization of natural remanent magnetization (NRM) and TRM to allow determination of the NRM/TRM ratio as a function of coercive force. This method requires a single heating above the Curie point of magnetic minerals to produce the full TRM. Laboratory alteration during TRM acquisition is monitored by comparing the ratio of ARM before (ARM_1) and after (ARM_2) heating. ARM is used as a substitute for TRM to diminish alteration arising from laboratory heatings such as those required in the Thellier and Thellier (1959) paleointensity method. Nevertheless, the heating above the Curie point in Shaw method may result in changes in ARM and TRM acquisition capacity. The slope (Kono, 1978) and individual TRM data correction (Rolph and Shaw, 1985) have been proposed using the ARM_1 and ARM_2 values at each AF demagnetization step.

The Shaw method has been widely used and it appears well suited for single domain (SD) and pseudo-single domain (PSD) magnetite ensembles (e.g., Tanaka, 1999). Volcanic rocks show complex mineralogy and a wide range of grain sizes. Further, magnetochemical changes due to heating may occur at different temperatures below and above the Curie point, which present a difficult challenge for paleointensity determination. In this study, we selected 49 samples from six Late Quaternary basaltic lava flows from central Mexico because of their low magnetic viscosity index, stable remanent magnetization, and close to reversible continuous thermomagnetic curves. Forty-two samples previously yielded high quality Thellier paleointensity results (Morales et al., 2001). We applied Shaw's (1974) original method and all three modifications (Kono, 1978; Rolph and Shaw, 1985; Tsunakawa and Shaw, 1994) on these carefully selected samples.

Magnetic Characteristics of Selected Samples

The volcanic lava flows selected for the present study are located in the southern part of the Basin of Mexico. All sites belong to *Sierra Chichinautzin* group, which represent youngest eruptive activity in the Trans-Mexican Volcanic Belt. The eruptive phase of this group probably occurred between Pleistocene and Holocene time (Urrutia-Fucugauchi and Martin del Pozzo, 1993). These monogenetic volcanoes, which outcrop around Mexico City, seem to be originated by subduction of Cocos plate under southern Mexico. Radiometric dates (using ^{14}C and K-Ar methods) are available for four out of six studied sites. Only relative ages are known for the two other volcanic lava flows. These former ages are based on geological and morphologic observations and previous paleomagnetic studies (Urrutia-Fucugauchi and Martin del Pozzo, 1993). Studied samples cover an age interval from about 0.39 My to 2000 years approximately. The detailed age information and location for each site may be found in Morales et al. (2001).

Magnetic characteristics of typical samples selected for Shaw paleointensity measurements are summarized in Figure 1 and could be described as follow: 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 (550-580 °C). The cooling and heating curves are reasonably reversible, which attest to the high magnetic stability of selected samples as also revealed by room-temperature susceptibility measurements (Figure 1a). Polished section observations under reflected light (Figure 1c) show the presence of a titanomagnetite phase associated with exsolved ilmenite of trellis texture.

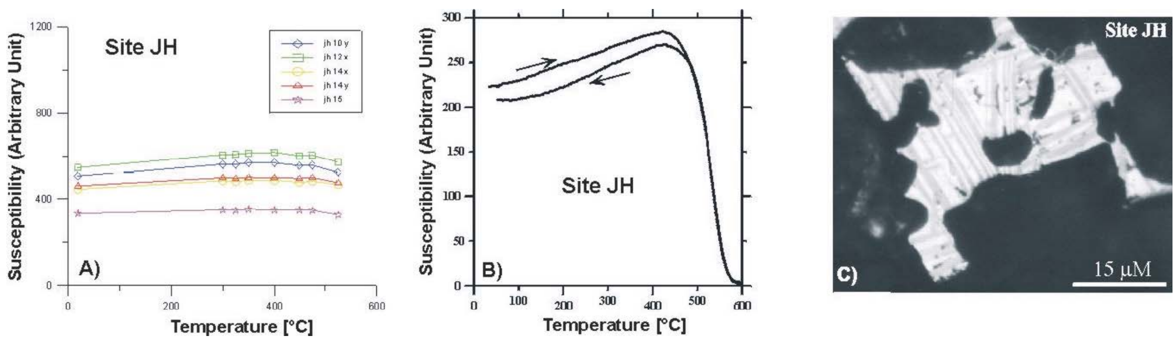

Figure 1. a) Representative curves of low field susceptibility against temperature (measurements were recorded at room temperature), b) Continuos susceptibility measurement (k-T curve) recorded on selected sample. The arrows indicate the heating and cooling curves, c) Representative reflected light microphotograph of studied Basalt, oil immersion, crossed nicols. Example of ilmenite of the 'trellis' type in titanomagnetite transformed into Ti-poor titanomagnetite.

Natural remanent magnetization carried by these grains should be a TRM, because this paragenesis typically develops at temperatures higher than the Curie point of magnetite (~ 578°C), Haggerty, 1976. All samples carry essentially a single component of remanent magnetization, observed upon both AF and thermal treatment (Morales et al., 2001). The median destructive fields (MDF) range mostly from 50 to 60 mT, suggesting 'small' pseudo-single domain grains as remanence carriers. Alternatively this may point to mixture of multidomain (MD) and a significant amount of SD grains.

Paleointensity Experiments and Results

We used the same experimental procedures as described in Shaw (1974) and Tsunakawa and Shaw (1994). All the specimens have been heated twice at 600°C in a 50 μT magnetic field in order to impart a full TRM. ARMs are given in a DC field of 50 μT at a maximum AF peak value of 100 mT. AF demagnetization was carried out up to 100 mT. Most of NRMs lose more than 90% of their original intensity after AF demagnetization.

With the data set obtained the graphs of ARM_1 vs ARM_2, NRM vs TRM and NRM vs TRM* [Rolph and Shaw's (1985) modification – TRM* = TRM x (ARM_1/ ARM_2)] were elaborated for each specimen, using the peak AF value as a parameter (Figure 2 and 3; Table 1). Moreover, following the approach of Tsunakawa and Shaw (1994), a second TRM (we call here TRM_2) and a third ARM (ARM_3) were applied to all the specimens as a further check for the validity of the Shaw method. Corresponding graphs: TRM vs TRM_2, ARM_2 vs ARM_3 and TRM vs TRM_2* [TRM_2* = TRM_2 x (ARM_2/ ARM_3)] were also plotted for each specimen (Figure 2 and 3; Table 1).

Thirty-six out of 49 samples failed to give any result by original Shaw method (Table 1) since their corresponding ARM_1 vs ARM_2 slopes were different from unity (the difference was more than 5% for rejected samples). Only 3 samples (JD2X, JD2Y, JE8X), however, satisfy the validity test proposed by Tsunakawa and Shaw (1994). Laboratory field estimates deduced from TRM vs TRM_2 are significantly higher/lower than expected (50 μT) for remaining samples (Table 1). Same is true for whole data set obtained by Kono's (1978) and Rolph and Shaw's (1985) corrections. Only exception is sample JM4, which yielded correct laboratory field estimate within 5% error using Rolph and Shaw's (1985) correction (Table 1).

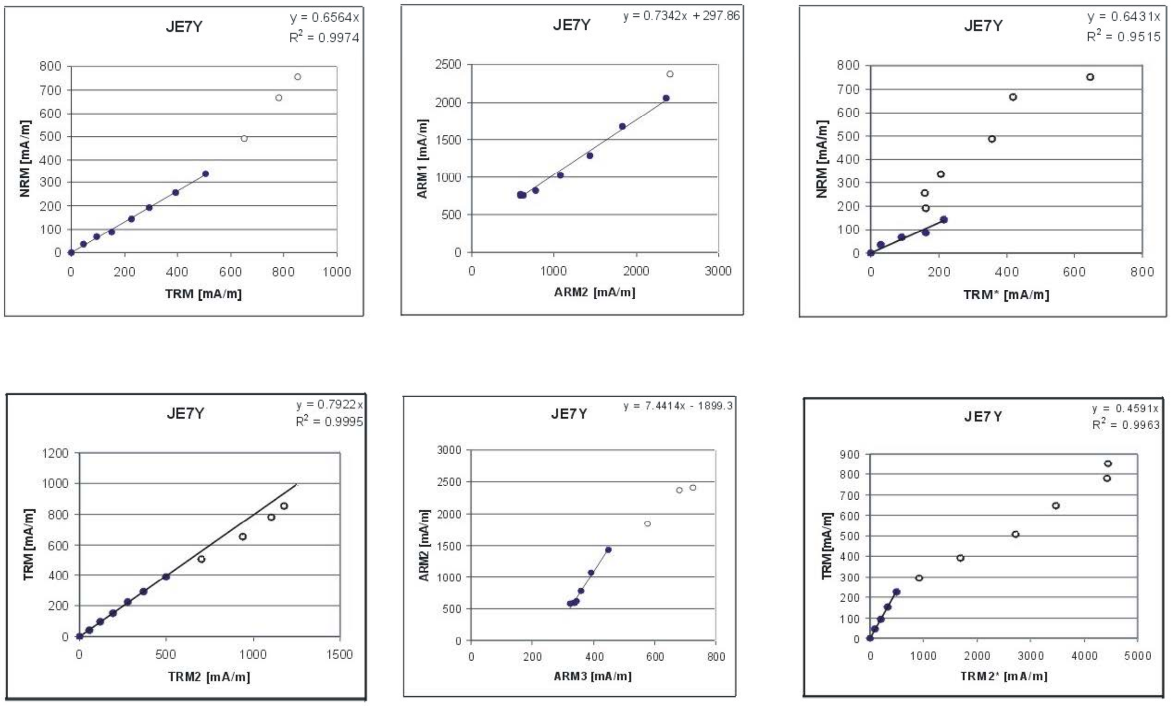

Figure 2. NRM-TRM, ARM_1-ARM_2 , NRM-TRM*, TRM-TRM_2, ARM_2-ARM_3 and TRM-TRM_2* graphs for representative sample. See text and Shaw (1974) for explanation.

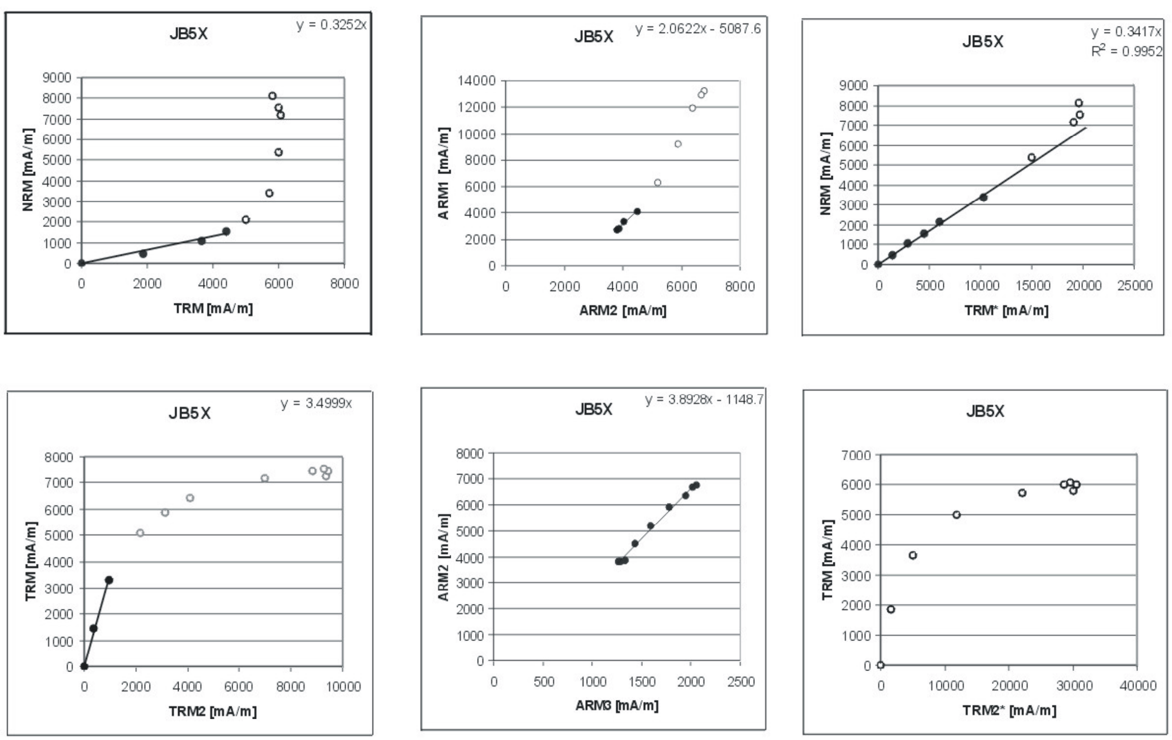

Figure 3. NRM-TRM, ARM_1-ARM_2 , NRM-TRM*, TRM-TRM_2, ARM_2-ARM_3 and TRM-TRM_2* graphs for representative sample. See text and Shaw (1974) for explanation.

					Shaw, 74	Kono, 78			RS, 85
Sample	**ARM2-ARM1**	**n/N**	**TRM-NRM**	**n/N**	**PI**	**PI**	**TRM*-NRM**	**n/N**	**PI**
JB5X	0.48	5 of 10	0.32	5 of 11	NR	7.7	0.38	9 of 11	18.9
JB5Y	0.39	7 of 13	0.33	4 of 10	NR	6.4	0.38	9 of 11	18.9
JB6X	0.57	13 of 13	0.32	5 of 11	NR	9.2	0.52	5 of 11	26.0
JB6Y	0.52	13 of 13	0.31	5 of 11	NR	8.1	0.42	9 of 11	20.9
JB7X	0.61	13 of 13	0.30	5 of 11	NR	9.2	0.51	5 of 10	25.3
JB7Y	0.53	13 of 13	0.37	5 of 11	NR	9.8	0.48	5 of 10	24.0
JB8X	0.67	13 of 13	0.43	5 of 11	NR	14.5	0.56	5 of 10	28.1
JB8Y	0.66	10 of 13	1.00	5 of 11	NR	33.1	0.30	5 of 10	14.9
				Mean =		**12.2**		**Mean =**	**22.1**
				σ =		**8.8**		**σ =**	**4.4**
JD2X	1.04	9 of 14	0.38	8 of 14	19.1	19.9	0.48	13 of 14	24.2
JD2Y	1.06	9 of 14	0.38	8 of 14	18.8	19.9	0.48	13 of 14	23.9
JD9X	1.31	9 of 14	0.40	7 of 14	NR	26.3	0.58	14 of 14	29.2
JD9Y	1.57	8 of 14	0.44	9 of 14	NR	34.4	0.68	14 of 14	33.8
JD10X	0.69	7 of 14	0.47	9 of 14	NR	16.2	0.52	14 of 14	26.2
JD10Y	1.04	9 of 14	0.51	10 of 14	25.4	26.5	0.54	12 of 14	26.9
JD11X	1.02	9 of 14	0.59	4 of 14	29.3	30.0	0.61	5 of 14	30.6
JD11Y	1.67	7 of 14	0.47	7 of 14	NR	39.0	0.54	5 of 14	26.8
				Mean =	**23.2**	**26.5**		**Mean =**	**27.7**
				σ =	**5.1**	**7.8**		**σ =**	**3.4**
JE4X	1.04	5 of 11	1.30	5 of 11	64.9	67.4	1.14	5 of 11	57.0
JE4Y	1.04	5 of 11	1.28	5 of 11	64.1	66.8	0.57	5 of 11	28.4
JE5X	1.11	9 of 11	0.61	8 of 11	NR	33.8	0.45	5 of 11	22.3
JE5Y	1.20	10 of 11	0.71	8 of 11	NR	42.7	0.64	5 of 11	31.8
JE7X	1.81	8 of 11	0.65	8 of 11	NR	58.9	0.72	5 of 11	36.1
JE7Y	1.35	10 of 11	0.66	8 of 11	NR	44.4	0.64	5 of 11	32.2
JE8X	0.97	5 of 11	0.53	8 of 11	26.4	25.5	0.54	4 of 11	26.8
JE8Y	1.29	8 of 11	0.56	8 of 11	NR	35.8	0.46	5 of 11	22.9
				Mean =	**51.8**	**46.9**		**Mean =**	**32.2**
				σ =	**22.0**	**15.7**		**σ =**	**11.1**
JH2X	0.68	6 of 14	0.47	4 of 14	NR	16.2	NR		NR
JH2Y	1.01	6 of 14	0.41	4 of 14	20.3	20.4	NR		NR
JH3	1.01	7 of 14	0.38	7 of 14	18.8	19.0	NR		NR
JH9X	1.09	6 of 14	0.45	4 of 14	NR	24.4	NR		NR
JH9Y	1.08	6 of 14	0.46	4 of 14	NR	24.7	NR		NR
JH12X					NR	0.0	NR		NR
JH13X	1.17	6 of 14	0.41	6 of 14	NR	23.9	NR		NR
JH13Y	0.72	5 of 14	0.60	3 of 14	NR	21.8	NR		NR
				Mean =	**19.5**	**21.5**			
				σ =	**1.1**	**8.1**			
JJ4X	1.00	8 of 13	1.04	6 of 14	52.2	52.1	0.73	6 of 14	36.3
JJ4Y	0.89	8 of 13	0.69	4 of 14	NR	30.8	1.08	14 of 14	54.0
JJ5X	0.85	13 of 13	0.99	14 of 14	NR	42.1	0.84	14 of 14	41.9
JJ5Y	0.76	13 of 13	0.80	9 of 14	NR	30.7	0.73	14 of 14	36.4
JJ6	0.77	12 of 14	1.01	14 of 14	NR	38.8	0.83	14 of 14	41.7
JJ7X	1.11	14 of 14	0.77	10 of 14	NR	42.8	1.04	14 of 14	52.2
JJ7Y	1.20	14 of 14	0.64	5 of 14	NR	38.4	1.16	14 of 14	58.1
JJ7Z	1.19	12 of 14	0.77	7 of 14	NR	45.4	1.25	13 of 14	62.6
				Mean =	**52.2**	**40.1**		**Mean =**	**47.9**
				σ =		**7.2**		**σ =**	**10.1**
JM1Y	1.39	11 of 11	0.49	11 of 11	NR	34.2	0.25	4 of 11	12.3
JM2Y	1.44	11 of 11	0.42	11 of 11	NR	30.3	0.20	4 of 11	10.1
JM3Y	1.15	11 of 11	0.44	10 of 11	NR	25.5	0.32	4 of 11	15.9
JM4	1.70	11 of 11	0.48	10 of 11	NR	40.5	0.21	3 of 11	10.7
JM5Y	1.48	11 of 11	0.67	11 of 11	NR	49.4	0.28	4 of 11	13.8
JM6Y	1.31	11 of 11	0.55	11 of 11	NR	35.7	0.21	4 of 11	10.3
JM7Y	1.06	10 of 11	0.70	11 of 11	35.0	37.3	0.46	5 of 11	23.2
JM8Y	1.04	11 of 11	1.02	11 of 11	51.1	53.0	0.66	5 of 11	33.1
JM9Y	1.06	11 of 11	0.96	11 of 11	47.9	50.8	0.55	4 of 11	27.6

Table 1. Paleointensity determination using Shaw's original (1974) and modified (Kono, 1978; Rolph and Shaw, 1985; Tsunakawa and Shaw, 1994) methods on Mexican late Quaternary basalts: n is number of NRM-TRM points used for determination, N is total number of available points. PI is the individual paleointensity estimate. $H_{lab\text{-}est}$ is a field value estimated from $TRM\text{-}TRM_2$ plot (Tsunakawa and Shaw, 1994). See text, Shaw, 1974 and Tsunakawa and Shaw, 1994 for explanation of TRMs and ARMs. RS, 85 – Rolph and Shaw, 1985; TS, 94 – Tsunakawa and Shaw, 1994.

Table 1(Continued)

					TS, 94 Shaw, 74	TS, 94 Kono, 78			TS, 94 RS, 85
Sample	**ARM3-ARM2**	**n/N**	**TRM-TRM2**	**n/N**	**$H_{lab-est}$**	**$H_{lab-est}$**	**TRM2*-TRM**	**n/N**	**$H_{lab-est}$**
JB5X	0.255	9 of 9	3.40	3 of 11	170.2	43.4	0.7043	3 of 11	35.2
JB5Y	0.216	9 of 9	3.76	3 of 11	188.2	40.7	0.7185	3 of 11	35.9
JB6X	0.060	6 of 9	1.78	3 of 11	89.1	5.3	0.2040	3 of 11	10.2
JB6Y	0.072	9 of 9	0.87	4 of 11	43.6	3.1	0.1800	3 of 11	9.0
JB7X	0.060	9 of 9	1.68	3 of 11	84.0	5.0	0.1719	3 of 11	8.6
JB7Y	0.065	6 of 9	1.9	3 of 11	92.8	6.0	0.1817	3 of 11	9.1
JB8X	0.122	9 of 9	1.72	3 of 11	86.0	10.5	0.2487	3 of 11	12.4
JB8Y	0.136	9 of 9	1.78	3 of 11	88.9	12.1	0.1154	3 of 11	5.8
				Mean =	**105.3**	**15.8**		**Mean =**	**15.8**
				σ =	**48.4**	**16.5**		**σ =**	**12.4**
JD2X	0.178	9 of 13	0.96	7 of 14	48.1	8.6	0.499	5 of 14	25.0
JD2Y	0.214	10 of 13	0.98	7 of 14	49.1	10.5	0.442	6 of 14	22.1
JD9X	0.199	11 of 13	0.83	7 of 14	41.7	8.3	0.495	5 of14	24.8
JD9Y	0.185	10 of 13	0.72	7 of 14	36.0	6.7	0.3945	5 of 14	19.7
JD10X	0.184	9 of 13	0.78	7 of 14	38.8	7.1	0.362	5 of 14	18.1
JD10Y	0.197	9 of 13	0.74	7 of 14	36.9	7.3	0.369	5 of 14	18.5
JD11X	0.206	9 of 13	0.78	7 of 14	39.1	8.1	0.371	5 of 14	18.6
JD11Y	0.177	8 of 13	0.88	7 of 14	44.1	7.8	0.363	5 of 14	18.2
				Mean =	**41.7**	**8.0**		**Mean =**	**20.6**
				σ =	**5.0**	**1.2**		**σ =**	**2.9**
JE4X	0.117	8 of 11	0.62	11 of 11	30.9	3.6	0.389	4 of 11	19.5
JE4Y	0.146	8 of 11	0.62	11 of 11	31.1	4.5	0.354	5 of 11	17.7
JE5X	0.150	8 of 11	0.84	8 of 11	42.0	6.3	0.467	5 of 11	23.3
JE5Y	0.137	8 of 11	0.67	11 of 11	33.7	4.6	0.356	5 of 11	17.8
JE7X	0.099	8 of 11	0.70	11 of 11	34.9	3.5	0.384	5 of 11	19.2
JE7Y	0.132	8 of 11	0.70	11 of 11	35.0	4.6	0.445	5 of 11	22.2
JE8X	0.279	9 of 11	0.94	8 of 11	46.8	13.1	0.724	5 of 11	36.2
JE8Y	0.143	8 of 11	1.16	7 of 11	57.8	8.3	0.515	5 of 11	25.8
				Mean =	**39.0**	**6.1**		**Mean =**	**22.7**
				σ =	**9.4**	**3.2**		**σ =**	**6.1**
JH2X	0.112	7 of 14	1.62	4 of 13	80.9	9.1	0.1745	5 of 13	8.7
JH2Y	0.117	7 of 14	1.20	6 of 13	59.8	7.0	0.2035	11 of 13	10.2
JH3	0.134	7 of 14	1.15	6 of 13	57.6	7.7	0.2021	11 of 13	10.1
JH9X	0.111	7 of 14	1.01	5 of 13	50.4	5.6	0.1773	10 of 13	8.9
JH9Y	0.111	7 of 14	1.03	5 of 13	51.7	5.7	0.1695	11 of 13	8.5
JH12X			NR						0.0
JH13X	0.146	7 of 14	1.14	4 of 13	57.1	8.3	0.1905	11 of 13	9.5
JH13Y	0.105	7 of 14	1.17	4 of 13	58.5	6.1	0.1849	11 of 13	9.2
				Mean =	**59.4**	**7.1**		**Mean =**	**9.3**
				σ =	**10.1**	**1.3**		**σ =**	**3.3**
JJ4X	0.130	10 of 14	0.87	7 of 13	43.7	5.7	0.3634	7 of 14	18.2
JJ4Y	0.094	9 of 14	0.90	5 of 13	45.2	4.3	0.4224	4 of 14	21.1
JJ5X	0.065	9 of 14	0.77	13 of 13	38.6	2.5	0.3418	7 of 14	17.1
JJ5Y	0.061	5 of 14	0.88	7 of 13	43.9	2.7	0.3896	5 of 14	19.5
JJ6			0.73	13 of 13	36.5	0.0			
JJ7X	0.063	10 of 14	0.73	13 of 13	36.5	2.3	0.3177	6 of 14	15.9
JJ7Y	0.066	10 of 14	0.98	7 of 13	49.2	3.2	0.4429	6 of 14	22.1
JJ7Z	0.107	12 of 14	0.91	8 of13	45.4	4.9	0.4706	5 of 14	23.5
				Mean =	**42.4**	**3.2**		**Mean =**	**19.6**
				σ =	**4.7**	**1.8**		**σ =**	**2.8**
JM1Y	0.064	8 of 11	0.52	6 of 11	26.2	1.7	0.6138	4 of 11	30.7
JM2Y	0.078	7 of 11	0.70	6 of 11	34.9	2.7	0.6982	4 of 11	34.9
JM3Y	0.065	5 of 11	1.02	5 of 11	50.8	3.3	0.781	4 of 11	39.1
JM4	0.023	6 of 11	1.32	6 of 11	66.0	1.5	1.0946	4 of 11	54.7
JM5Y	0.049	6 of 11	0.76	6 of 11	38.0	1.9	0.7302	4 of 11	36.5
JM6Y	0.077	6 of 11	0.74	6 of 11	36.9	2.8	0.7627	4 of 11	38.1
JM7Y	0.124	7 of 11	0.82	6 of 11	40.8	5.1	0.5522	4 of 11	27.6
JM8Y	0.040	6 of 11	3.96	4 of 11	198.1	7.9	0.7839	4 of 11	39.2

Discussion

The Shaw method and its variants are routinely used for paleointensity determination. However, both the original and the modified methods occasionally give unreasonable paleointensities (Sherwood et al., 1988; Goguitchaichvili et al., 1999) and it is still unclear whether ARM corrections are capable of offsetting satisfactorily all kinds of TRM capacity changes. As showed by Kono (1987), changes in ARM spectra do not always completely follow TRM spectra changes after intensive heating. Tsunakawa and Shaw (1994) modification of the Shaw's original method uses a double heating of samples above the Curie point. After heating the specimen twice in the same field, the measured intensity from the second heating is compared with the laboratory field intensity. If the difference is larger than the experimental error, then it is assumed that the ARM correction is not applicable. Using this selection technique Tsunakawa and Shaw (1994) obtained 40 per cent success in determining the geomagnetic paleointensity from historic lava flows. The experiments were done however on limited number of samples and they reliability should be checked by more robust data-set.

More than 50% of samples of Tsunakawa and Shaw's (1994) collection are characterized by relatively low Curies temperatures ranging from 280 to 360°C, which may contain thermally unstable 'large PSD' or dominantly MD magnetic grains, which are rather unsuitable material for absolute paleointensity experiments. TRM and pTRMs in MD grains present wide spectra of unblocking temperatures, being partly demagnetized from room temperature to the Curie point. Recent experiments by Dunlop and Ozdemir (2001) on narrow-band pTRMs show that stepwise thermal demagnetization has a sharp spectrum for SD, a broad Gaussian spectrum for MD, and non-Gaussian spectra for PSD grains.

Shaw's paleointensity method and its modifications yield extremely low success rate in our experiments to determine the absolute intensity of geomagnetic field. In contrast, Thellier and Thellier (1959) method applied on samples from the same lava flows results in a high success rate with 42 individual reliable determinations (Morales et al., 2001). The NRM fractions used for paleointensity determination range from 0.34 to 0.97 and the quality factors (Coe et al., 1978) varies between 4.5 and 97.8, being normally greater than 5 which indicates to high technical quality data. Almost 95% failure obtained using Shaw's method may be mainly because the ARM correction is only a first-order correction and that once thermal alteration becomes excessive the correction breaks down (Tsunakawa and Shaw, 1994). Moreover, if the TRM/ARM ratio changes after the laboratory heating, then the correction methods may become invalid. Magneto-chemical alteration of volcanic flows, in our case, seems to occur when heating the samples above the Curie temperature. The Thellier method uses data at lower temperatures before significant magnetochemical changes take place. For those cases, it seems to be more powerful method to determine geomagnetic paleointensities.

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An attempt to determine the microwave paleointensity on historic Paricutín volcano lava flows, Central Mexico

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RESUMEN

Reportamos un estudio preliminar de magnetismo de rocas y paleointensidad con la técnica de microondas de los flujos de lava históricos entre 1943 y 1948 del volcán Paricutín. Esos flujos de lava muestran afloramientos frescos, bien preservados y expuestos. La mayoría de las muestras estudiadas se caracterizan por diagramas ortogonales simples. Las curvas de magnetización remanente isotermal (IRM) muestran saturación en campos bajos a moderados, sugiriendo la serie de titanomagnetitas. De los experimentos de histéresis se determinó que los portadores magnéticos son probablemente titanomagnetitas ricas en fierro con un comportamiento de dominio sencillo a pseudo-sencillo. La técnica de paleointensidad por microondas se aplicó a tres muestras seleccionadas usando el método de Kono y Ueno (1977); i.e., la dirección del campo aplicado en laboratorio fue perpendicular a la dirección de la magnetización remanente. Los resultados de paleointensidad fueron de 11.39, 25.6 y 58.1 μT, que son significantemente diferentes de los valores esperados. La dispersión observada puede deberse a la pequeña fracción de la magnetización remanente natural usada para la determinación de la paleointensidad, o de no haber usado una corrección por la razón de enfriamiento para muestras naturales.

Palabras Clave: Paleointensidad, Técnica Microondas, Magnetismo de Rocas, México Central, Volcán Paricutín.

ABSTRACT

We report a preliminary rock-magnetic and microwave paleointensity study of historic lava flows between 1943 and 1948 of Paricutin volcano. These lava flows show well-preserved, well-exposed, fresh and extensive outcrops. Most samples are characterized by simple uni-vectorial plots. The isothermal remanent magnetization (IRM) curves show saturation at low to moderate fields, suggesting titanomagnetite series. From hysteresis experiments magnetic carriers are likely iron-rich titanomagnetites with single-domain or pseudo-single-domain behavior. Microwave paleointensity technique was applied to three selected samples using Kono and Ueno's (1977) method; i.e., the direction of applied laboratory field was perpendicular to the direction of remanent magnetization. The samples yielded paleointensity values of 11.39, 25.6 y 58.1 μT, which are scattered and significantly different from expected values. The observed scatter is likely to be caused by the small fraction of natural remanent magnetization used for paleointensity determination, or from not using a cooling rate correction for natural samples.

Key Words: Paleointensity, Microwave Technique, Rock-magnetism, Central Mexico, Paricutin Volcano.

Introduction

The intensity of a paleofield may be estimated by comparing the intensity of the thermoremanent magnetization (TRM) acquired by a volcanic rock at the time of emplacement with the strength of a laboratory TRM produced by a known field (Koenisberger, 1938). Heating of the rock above its maximum blocking temperature alters initial magnetic mineralogy and may disturb the relationship between field and TRM. Thellier and Thellier (1959) devised a method to study partial TRM (pTRM), obtained from parts of the blocking/unblocking temperature spectra not affected by magneto-mineralogical alteration. However, for most of natural rocks the alteration may occur at very low temperatures, which impedes an accurate determination of ancient geomagnetic field strength (Kosterov and Prévot, 1998).

Conventional TRM is formed when heat in the form of phonons is sufficient to generate spin waves (magnons) within the individual magnetic domains (Walton et al., 1993). The spin waves allow magnetization to reverse, and during cooling the magnetization becomes fixed with a statistical bias towards the ambient magnetic field direction. Alternatively, spin waves can be generated within magnetic grains by direct microwave excitation (Shaw et al., 1996). The TRM formed by this method is almost identical to the one formed by heating, except that the magnetic system is heated and cooled very quickly (a few seconds). Thus microwave TRM does not cause any noticeable alteration of the magnetic minerals (Walton et al., 1993). Microwave techniques seem to be ideal for determining paleointensities.

It is extremely valuable to study historic lava flows as the geomagnetic field at time of extrusion is known. In this study, we report a preliminary rock-magnetic and microwave paleointensity investigation of historic lava flows of Paricutin volcano, from the period between 1943 and 1948. These lava flows feature well preserved, well-exposed, fresh, extensive outcrops.

Paricutin Volcano

Paricutin Volcano emerged on February 20, 1943 in a corn field near San Juan Parangaricutiro village, Michoacan, Mexico (Ordoñez, 1943). This area is in the Michoacan-Guanajuato volcanic field, which is characterized by numerous cinder cones and medium sized shield volcanoes. In this area, Jorullo volcano was born on September 29, 1759, some 72 km to the southeast (Figure 1).

The birth and growth of Paricutin volcano have been extensively discussed Fries, (1953), Segerstrom, (1965), Luhr and Simkin, (1993) and others. Activity extended from 1943 trough 1952 (Figure 1). The composition and petrography of erupted material changed from olivine-bearing basaltic andesites in 1943 to orthopyroxene-bearing andesite at the end of 1952 (Wilcox, 1954).

Rock-Magnetic Properties of Paricutin Samples

The NRM measurements were performed using a Molspin spinner magnetometer. The low-field susceptibility was measured with a Bartington MS2 system equipped with dual frequency laboratory sensor. Stability and vectorial composition of natural remanence were investigated by stepwise alternating field (AF) demagnetization in 9 to 11 steps up to a maximum of 100 mT, using a Schonstedt demagnetizer in the three-axes stationary mode.

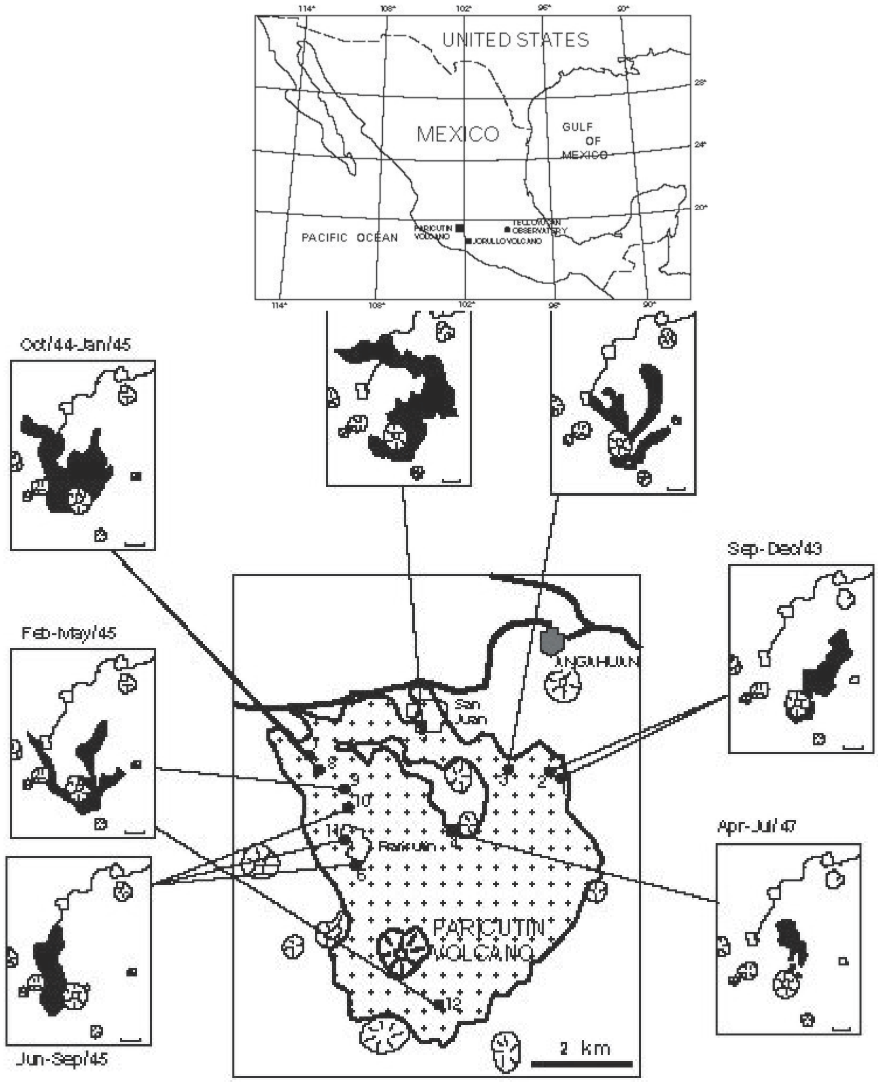

Figure 1. Schematic map of the Paricutin Volcano and its lava field (shaded area). The distribution of flows in terms of lava eruptive episodes is shown in the small maps, with the dates of the episode and the aeral extension of lavas (maps adopted from Luhr and Simkin, 1993).

The viscosity index (Thellier and Thellier, 1944; Prévot, 1983) may estimate the capacity of a sample to acquire a viscous remanent magnetization, and its paleomagnetic stability. We placed the samples during 10 days with one axis aligned with Earth's magnetic field. After measuring the magnetization (M_d), the samples were placed for another 10 days in a field-free space, and the magnetization (M_0) was measured again. The viscosity index is V = [(Z_d – Z_0) : M_{nrm}] x 100, where Z_d and Z_0 are the magnetization components of M_d and M_0 parallel to the magnetizing field, and M_{nrm} is the intensity of natural remanent magnetization. The values were lower than 2.7 %: thus, there is no evidence of significant viscous remanent magnetization in Paricutin lavas.

Most samples, including those used for paleointensity experiments are characterized by a simple uni-vectorial component pointing towards the origin (Figure 2a). The median destructive fields (MDF) range mostly from 30 to 40 mT, suggesting ‘small’ pseudo-single domain grains as remanent magnetization carriers (Dunlop and Özdemir, 1997).

Hysteresis measurements at room temperature were performed on all samples using an AGFM ‘Micromag’ apparatus at the paleomagnetic laboratory in Mexico City, with fields up to 1 Tesla. Saturation remanent magnetization (J_{rs}), saturation magnetization (J_s) and coercitive force (H_c) were calculated after correction for paramagnetic contribution. Coercivity of remanence (H_{cr}) was determined by applying progressively increasing backfield after saturation (Figure 2b and c). A representative hysteresis curve is shown in Figure 2 (b and c). The curves are symmetrical. Near the origin, no potbellied or wasp-waisted behaviors (Tauxe et al., 1996) were detected, which probably reflects restricted ranges of opaque mineral coercivities. Judging from the ratios of hysteresis parameters, all samples fall in the pseudo-single domain (PSD) grain size region (Day et al., 1977), probably indicating a mixture of multidomain (MD) with a significant amount of single domain (SD) grains. Corresponding isothermal remanence (IRM) acquisition curves were found to be very similar for all samples. Saturation is reached in moderate fields of the order of 150-200 mT, which points to some spinels as remanence carriers.

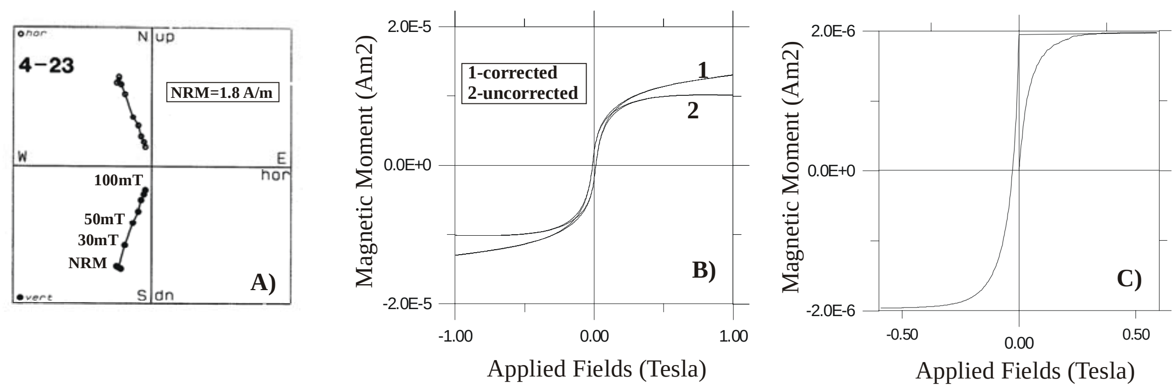

Figure 2. A) Typical example of alternating field demagnetization vector plot for Paricutin lavas. B) Example of hysteresis curve measured with 'AGFM-Micromag' system. C) Example of direct field isothermal remanent magnetization (IRM) acquisition curve and back-field curve.

Results and Discussion

The paleointensity experiments were carried out at the Geomagnetic Laboratory of the University of Liverpool, using the 8.2GHz Microwave system, under scientific supervision of Dr. John Shaw. The technique used, described in detail in Hill and Shaw (1999), is a variant of the stepwise Thellier method (Kono and Ueno, 1977). A microwave thermoremanence is induced perpendicular to the directions of NRM, so that a single microwave application is required for each power step (Figure 3 and 4). This eliminates the need for accurate reproducibility of microwave power (up to 140 W) absorbed by the sample. Control heatings (so-called pTRM checks) are normally used with the conventional Thellier type technique (Prévot et al., 1985) to monitor alteration during the experiment. In the case of microwave paleointensity it is difficult to repeat a previous step because the power absorbed by a rock sample is the one that needs to be reproduced.

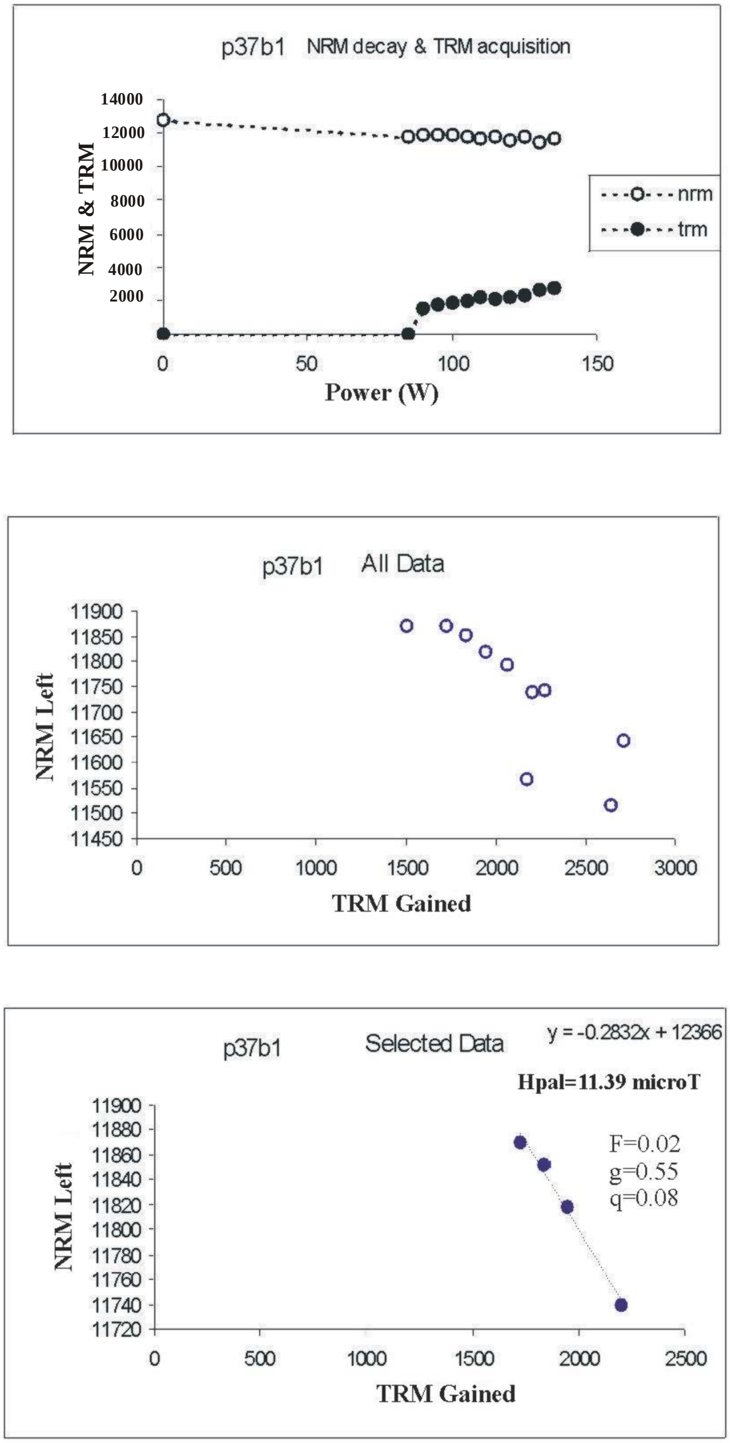

Figure 3 and 4. Results from the microwave intensity analyses: *f*, *g* and *q* are the fraction of extrapolated NRM used, the gap factor and quality factor (Coe et al., 1978) respectively. Open/close symbols devote to the points rejected/used for paleointensity determination.

Another way to check for alteration is to monitor hysteresis parameters (Hill and Shaw, 2000). However, Goguitchaichvili et al., (2001) showed that room temperature hysteresis parameters, in terms of the plot of magnetization ratio vs. coercivity ratio, lack resolution for most of natural rocks. No alteration test was performed in this study.

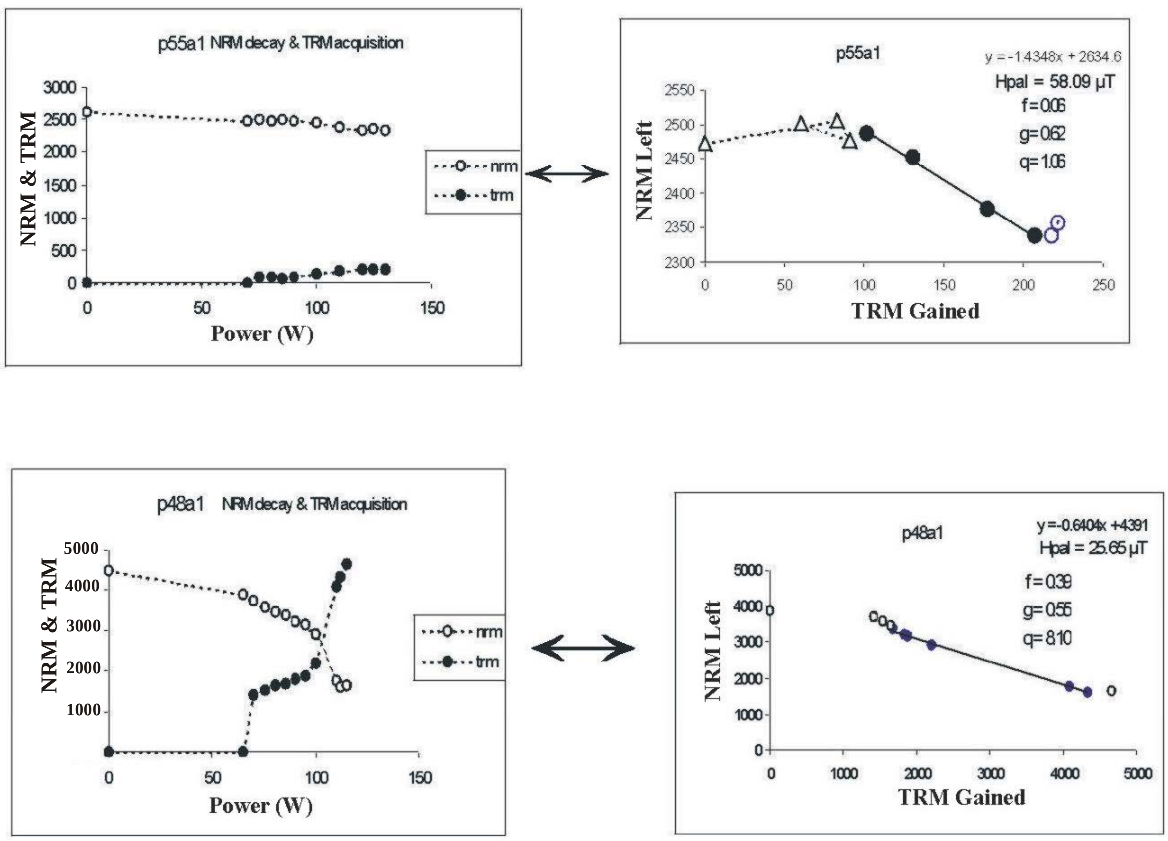

Figure 4

Three magnetically stable representative samples were selected for pilot microwave paleointensity experiments. Paleointensity data are reported on the Arai-Nagata (Nagata at al., 1963) plot on Figure 3 and 4. The samples yielded paleointensity values of 11.39, 25.6 y 58.1 μT, which are significantly different than the expected intensity of around 43 microTesla, from Teoloyucan Observatory data. Previous studies of Paricutin volcano

(Gonzalez et al. (1997) show similar scatter, suggesting that internal characteristics of the lava flows may be responsible for this fact. The failure of microwave paleointensity measurements on Paricutin samples is probably due to the fact that only a small fraction ($0.02 < f < 0.39$) of natural remanent magnetization was used for paleointensity determination. Additional measurements, using 14GHz sysytem are strongly needed. Another explanation may be the impossibility to use the cooling rate correction for natural samples. The samples are 'cooled' down in a few seconds during the microwave paleointensity experiment. Based on experimental studies Fox and Aitken (1980), McClellend, (1984), Goullpeau et al. (1989), Chauvin (1989), Aitken et al. (1991), Biquand (1994) and Garcia (1996) showed that there is no clear relation between the cooling rate and TRM acquisition. The differences between TRM produced with low and high cooling rates may vary between 5 and 600%. Bowles et al. (2002), studying some archeomagnetic material from Ecuador, demonstrated that the differences between cooling rates in the past and in laboratory may produce a large scatter of paleointensity data. However, microwave paleointensity analysis of the 1960 Kilauea flow (Hill & Shaw, 2000), and the experimental study of Hill et al. (2002) show that the field intensity can be correctly estimated without using cooling corrections. Additional historic lava flows need to be studied in order to ascertain whether the failure of paleointensity experiments is due to internal characteristic of Paricutin Volcano lava flows.

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Geomagnetic field strength during Late Miocene: first paleointensity results from Baja California

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Abstract. Thellier paleointensity experiments were carried out on 73 selected samples from the late Miocene Baja California volcanic flows for the first time. The selected samples showed a stable remanent magnetization observed upon thermal treatment and reasonably reversible continuous susceptibility curves, compatible with Ti-poor titanomagnetites. Judging from the ratios of hysteresis parameters, it seems that the samples fall in the small pseudo-single-domain grain size region. Thirty-three samples from 10 individual flows yielded acceptable paleointensity estimates. The NRM (natural remanent magnetization) fractions used for paleointensity determination range from 37 to 81% and the quality factor varies between 4.5 and 24.5, being normally greater than 5. The mean paleointensity values per flow obtained in this study range from 18.1 ± 3.1 to 45.6 ± 4.3 μT and the corresponding Virtual Dipole Moments (VDMs) range from 3.8 ± 0.6 to 8.8 ± 0.8 (10^{22} Am^2). These values correspond to mean value of 6.6 ± 1.7 x 10^{22} Am^2, which is about 85% of the present geomagnetic axial dipole. The average VDM value for late Miocene, according to all available, selected data, is 6.1 ± 1.5 (10^{22} Am^2), which is higher than average paleofield for the period 5-160 Ma [4.2 ± 2.3 x 10^{22} Am^2 after *Juarez and Tauxe*, 2000]. Given the large dispersion and the very poor distribution of reliable absolute intensity data, it is hard to draw any firm conclusions regarding the time evolution of the geomagnetic field. More high quality determinations are strongly needed.

1. Introduction

Coe et al. [2000] recently suggested that absolute intensity is a fundamental constraint in numerical models that promise to provide unprecedented insight into the operation of the geodynamo. Despite recent growth of the quality and quantity of absolute geomagnetic intensity measurements [*Brassart et al*., 1997; *Juarez et al*., 1998; *Carlut et al*., 1999; *Valet et al.*, 1999; *Juarez and Tauxe*, 2000; *Zhu et al*., 2000; *Carlut and Quidelleur*, 2000; *Valet and Herrero-Bervera*, 2000; *Goguitchaichvili et al*., 2001a,b] reliable absolute paleointensity data are still scarce and cannot be used yet to document the general characteristics of the Earth's magnetic field strength [*Selkin and Tauxe*, 2000; *Riisager et al*., 2002]. This is probably derived from the fact that reliable paleointensity values are generally much more difficult to obtain than reliable directional data, because only volcanic rocks satisfying some very specific magnetic criteria [*Calvo et al*., 2002, *Kosterov and Prévot*, 1998] can be used for absolute paleointensity determinations.

Judging from the existing paleointensity database [ftp://ftp.dstu.univ-montp2.fr/pub/paleointb, see also *Perrin et al.*, 1998] only large-scale features of geomagnetic paleointensity can be resolved, such as the pronounced lowering of the Earth's Virtual Dipole Moment (VDM) in the whole Mesozoic and part of Paleozoic [*Prévot et al.*, 1990], during which the dipole structure of the field was preserved [*Perrin and Shcherbakov*, 1997]. However, high geomagnetic intensities of (12.5 ± 1.4) x 10^{22} Am^2 have been recently reported [*Tarduno et al*., 2001] for the interval 113 – 116 Ma, which are consistent with some inferences from computer simulations of the Earth's dynamo [*Glatzmaier et al*., 1999]. *Zhu et al*. [2001] found relatively low dipole-field intensity just prior to the Cretaceous normal superchron. It appears, then, that the 'Mesozoic Dipole Low' is not firmly established yet.

Although some new paleointensity determinations have been reported during last years, Miocene paleointensity data are still scarce and of dissimilar quality [*Goguitchaichvili et al*., 2000a; *Riisager et al*., 2000], and therefore new reliable data are needed. In this paper we present the first paleointensity results obtained for Baja California volcanic rocks erupted mainly between 11 and 8 Ma, which have several advantages: (1) they are widely distributed in a large volcanic province of easy access; (2) they record faithfully the magnetic field that existed at the time of their eruption; (3) most of them are fresh for isotopic dating and have already yielded reliable K-Ar and Ar-Ar ages.

2. Analyses of Miocene Paleointensity Data

There is now a general agreement among the paleomagnetic community that the *Thellier and Thellier* [1959] method yields the most reliable paleoinetensity determination [*Goguitchaichvili et al*., 1999a; *Morales et al*., 2002]. Thus, in the following analysis, we will consider only determinations obtained with Thellier's technique.

It is very hard to assess the reliability of the worldwide paleointensity results, due to the poor documentation of many previously published determinations. In our analyses of the Miocene paleointensity data, we considered only results 1) obtained with the Thellier method for which 2) positive pTRM checks attest the absence of alteration during heatings, 3) at least 3 determinations exist per unit, 4) an error of the mean paleointensity about 20% or less, and 5) no data were considered from transitional polarity units [*Perrin and Shcherbakov, 1997*]. Moreover, we applied the same acceptance criteria on individual determinations as in the present study (see below). From more than 200 Miocene determinations available in the IAGA database [compiled by *Perrin and Schnepp*, see also *Perrin et al*., 1998] only 10 determinations from Steens Mountain [*Prévot et al*., 1985], 3 determinations from the Velay [French Massif Central, *Riisager et al.*, 2000], 2 determinations from submarine basaltic glasses (SBG) [*Juarez et al*., 1998] and 4 determinations from Central Mexico [*Goguitchaichvili et al.*, 2000a] fulfil these criteria (Figure 1).

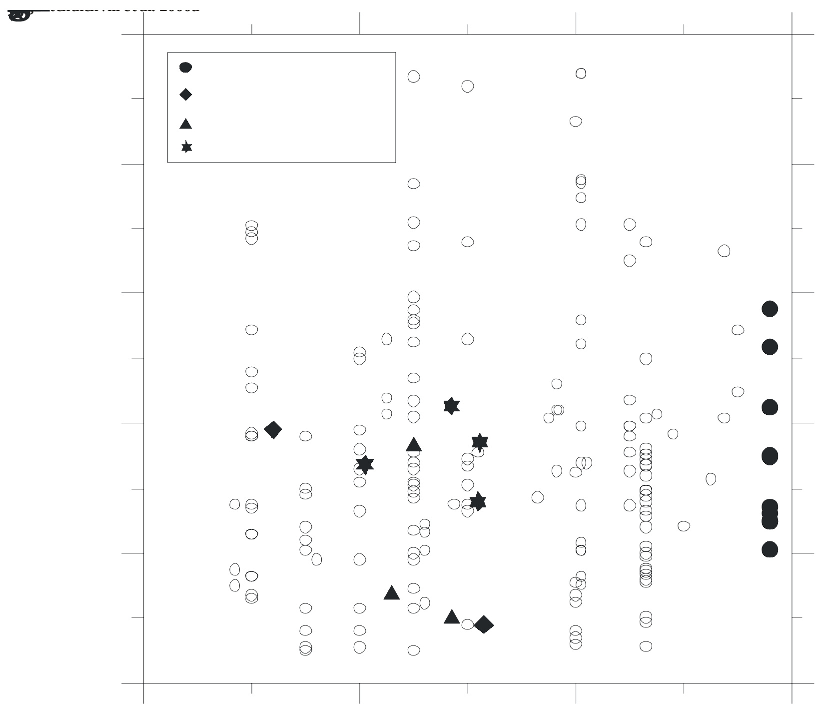

Figure 1. Virtual dipole moments against ages of volcanic units in My. Determinations fulfilling selection criteria are shown by filled symbols. Data are taken from Montpellier paleointensity database [ftp://ftp.dstu.univ-montp2.fr/pub/paleointb, see also *Perrin et al.*, 1998].

Even using these high standards in data selection, it is observed that there is a very high dispersion of paleointensity data between 16 and 6 My. Such dispersion may be due to experimental artefacts rather than reflecting a true feature of the geomagnetic field at that time. This point will be discussed later in more details.

Another feature of paleointensity database is that there are very few paleointensity determinations available for the period 16-20 My. *Dunlop and Hale* [1976] reported one determination and *Koenisberger* [1938] gave two preliminary determinations using a total TRM (thermoremanent magnetization) method. These results are very close to the present day field intensity. However, the first study is based on submarine basalts drilled near the Mid-Atlantic ridge, which were suspected to be maghemitized and the second work presents only a 'historical' interest, and consequently these were not included in our figure. Relatively low paleointensity values were reported from Central Chile between 19 – 17 My [*Goguitchaichvili et al*., 2000b]. Although a linear relationship was observed between the NRM (natural remanent magnetization) and TRM acquired during the Thellier experiments, undetected partial chemical alteration of the primary magnetizations may cause a reduction of the determined paleointensity values [*Goguitchaichvili et al*., 2000b]. Thus, more reliable absolute geomagnetic intensity determinations from Miocene rocks are strongly needed.

3. Sampling Localities and Ages

Vast areas of the northern and central parts of Baja California Sur are covered by Cenozoic volcanic rocks, which were studied by several authors [e.g. *Gastil et al*., 1979; *Hausback*, 1984; *Hagstrum et al*., 1987]. Based on petrography and geochemical composition, *Sawlan and Smith* [1984], and *Sawlan* [1991] identified two Mio-Pliocene lava suites on the northern part of Baja California Sur, apparently unrelated to each other. The first suite of lavas is tholeiitic and the second is alkaline. On a regional scale both types of lavas present general characteristics that, it was thought, should allow their easy identification. For instance, the tholeiitic lavas commonly form the cover of extensive mesas whereas the alkalic lavas were more commonly found in close association with cinder cones. Although all these relations were supported by the scarce information available at the time concerning the physical aspects of the lava flows, as more detailed studies were completed it became clear that the volcanism in the area of study presents a different character than reported by *Sawlan and Smith* [1984] and *Sawlan* [1991]. Indeed, *Cañón-Tapia and Rojas Beltrán* [2001] have shown that instead of having two independent suites of lava in the region, the volcanism of the zone marks the remains of one or two basaltic (sensu lato) volcanic fields, each formed by tens of eruptive centers. Both tholeiitic and alkaline volcanism were coeval on the region, and although it is true that most of the cinder cones documented by *Cañón-Tapia and Rojas Beltrán* [2001] presumably erupted alkaline lavas, a few vents were positively identified as the source of some of the tholeiitic products. Additionally, it was long time believed that volcanism at this area is related to the opening of the Gulf of California, but *Cañón-Tapia et al*., [2001] showed that intra-continental volcanism also seems to be involved; this type of volcanism occurred between the transition from the end of oceanic plate subduction at the West and the opening of the Gulf at the East.

As part of their efforts to study the details of the volcanic activity on the region, *Negrete-Aranda and Cañon-Tapia*, [2000] recently carried out a systematic paleomagnetic sampling at sites located in late Miocene lava flows in the area located between 27°30'N and 26°08'N latitude and between 112°20'W and 113°20'W longitude (Figure 2). Commonly, the outcrops in the region extend laterally over a few tens of meters. The samples were distributed throughout each flow both horizontally and vertically in order to minimize effects of block tilting and lightning. From their large collection (more than 300 standard paleomagnetic cores), we selected sites with the best directional behavior and available radiometric ages (Table 1) for paleointensity determination

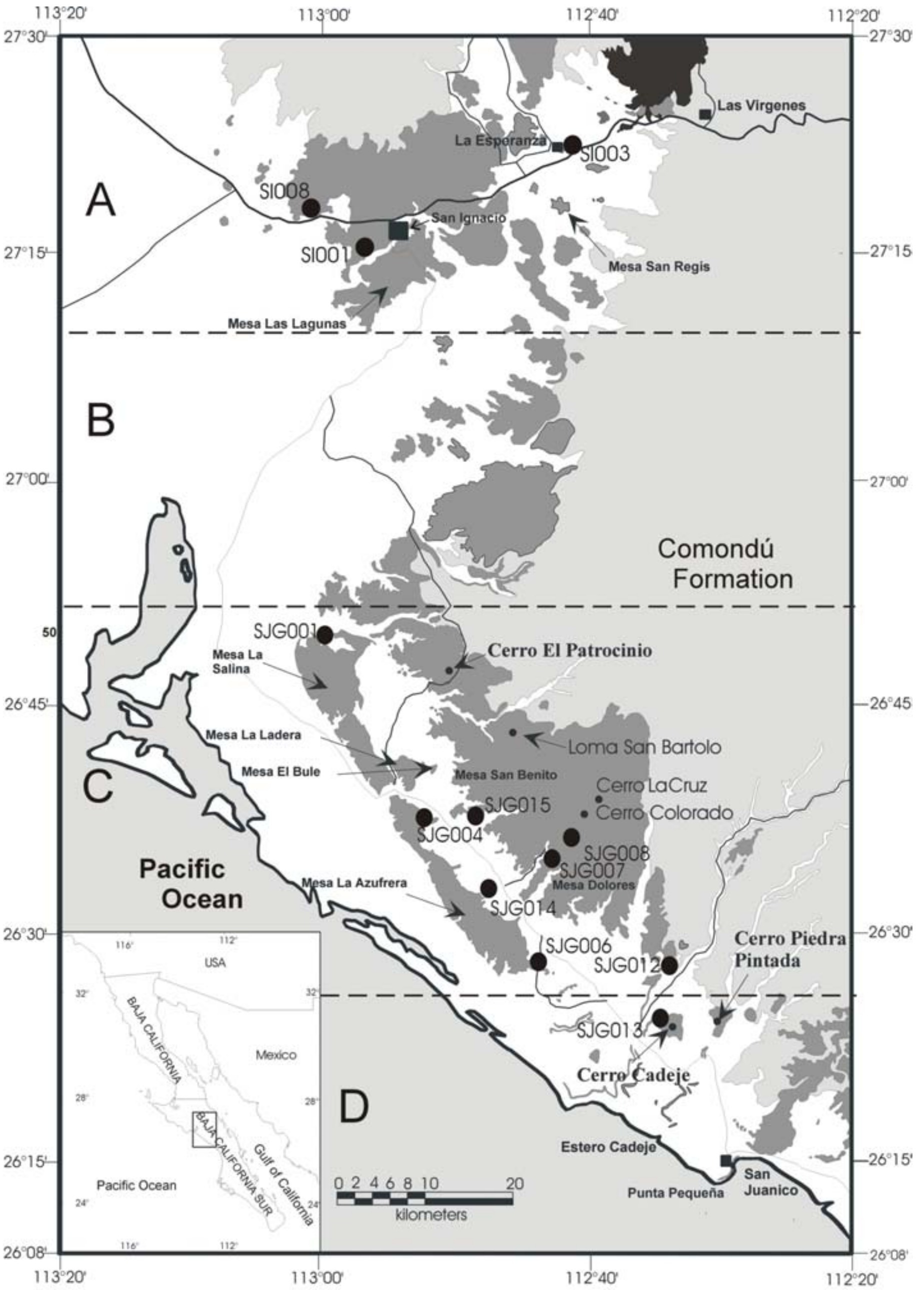

Figure 2. Map of the area of study showing location of the sites used to determine paleointensities. The shaded area shows the extent of the present day lava cover on the region that includes tholeiitic and alkaline lavas.

Table1. Coordinates and isotopic ages

Site	Lat. N	Long. W	Age (Ma)	Reference
SJG01	26°44.8'	112°59.8'	10.6 ± 0.4	*Aguillón-Robles et al.*, [2001]
SJG04	26°38.4'	112°51'8'	6.0 ± 2.6	*Rojas Beltrán*, [1999]
SJG06	26°28.2'	112°43.4'	10.1 ±1.0	*Rojas Beltrán*, [1999]
SJG07	26°34.0'	112°44.2'	11.7 ± 0.8	*Gastil et al.*, [1979]

Coordinates and isotopic ages of rocks within the area of study (see also text).

Ages of the studied units range from 11.7 to 6 My according to available radiometric dates (Table 1). Four sites (from, a total of ten sites selected for paleointensity experiments) are directly dated either by K-Ar or Ar-Ar methods (SJG01, SJG04, SJG06 and SJG07). Most of the remaining sites can be easily correlated with the dated sites, considering their chemical affinity and field observations [*Negrete-Aranda and Cañon-Tapia*, 2000]. Sites SI3, SJG12, 14 and 19 are all tholeiitic lava flows and most probably formed between 11 and 10 My. The chemical affinity of Sites SJG15 and SJG16 is unknown. However, based on field observations, they probably erupted between 8 and 11.7 My.

4. Rock-Magnetic Properties of Selected Samples

Four kinds of magnetic measurements, combined with microscopic observation under reflected light, were carried out in order to select the most suitable samples for Thellier paleointensity experiments. Progressive demagnetizations were carried out at both CICESE (Baja California) and National University of Mexico (UNAM) paleomagnetic laboratories. Rock-magnetic experiments, which included: viscosity index determination, temperature dependence of magnetic susceptibility and hysteresis experiments were performed at the paleomagnetic laboratory of UNAM. Figures 3, 4, 5 and 6 summarize magnetic properties of the typical samples used for paleointensity determination.

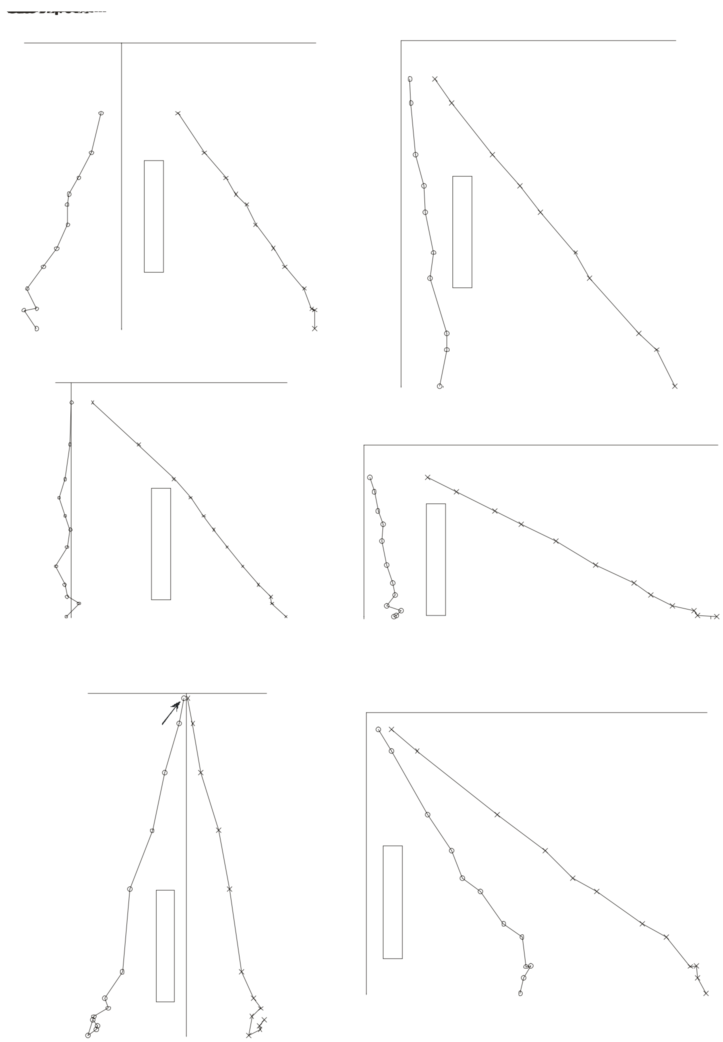

Figure 3. Orthogonal vector plots of stepwise thermal demagnetization (stratigraphic co-ordinates). The numbers refer to temperatures in °C. o – projections into the horizontal plane, x – projections into the vertical plane.

4.1. Viscosity Index

Determination of the viscosity index [*Thellier and Thellier*, 1944; *Prévot et al*. 1983] allows to estimate the capacity of a sample to acquire a viscous remanent magnetization, and is therefore useful to obtain information about their paleomagnetic stability. For this purpose, we placed the samples for 20 days with one of their axes aligned with Earth's magnetic field. After measuring their magnetization ($\mathbf{M_d}$), they were placed for another 20 days in a field-free space, and the magnetization (**M_0**) was measured again. This allows to calculate the viscosity index V = [(Z_d – Z_0) : M_{nrm}] x 100, where Z_d and Z_0 are respectively the magnetization components of M_d and M_0 which are parallel to the magnetizing field. **M_{nrm}** is the intensity of natural remanent magnetization. Although viscosity indexes varied between 0 and 22%, the values of selected samples for paleointensity determination were lower than 5%.

4.2. Stability of the Remanence

Selected samples carry essentially a single and stable component of magnetization, observed upon thermal treatment (Figure 3). Minor secondary components were easily removed applying 200 or 250°C of thermal demagnetization. The greater part of remanent magnetization, in most cases, was removed at temperatures between 400 and 550°C. This indicates to low-Ti titanomagnetites ($x < 0.2$) as responsible for remanent magnetization or a wide range grain size for the magnetic carriers.

4.3. Temperature Dependence of Magnetic Susceptibility and Microscopic Observation

Continuous low-field susceptibility measurements with temperature (χ(T) curves) were performed in air using a Bartington susceptibility meter MS2 equipped with furnace. One specimen per flow was heated up to 600°C at a heating rate of 10°C/min and then it was cooled at the same rate. In all cases, the Curie temperatures were determined by *Prévot et al*'s [1983] method. The curves show the presence of a single ferrimagnetic phase with Curie point compatible with Ti-poor titanomagnetite. The cooling and heating curves are reasonably reversible (Figure 4, upper part). In a few cases, we could not obtain a correct χ(T) curve because of the low magnetic signal of initial susceptibility.

The microscopic observations (under reflected light) on polished sections show that the main magnetic mineral is magnetite associated with exsolved ilmenite (Figure 4, bottom) probably formed as a result of oxidation of titanomagnetite during the initial flow cooling. These texture typically develops at temperatures higher than 600°C [*Haggerty*, 1976] and consequently, the natural remanent magnetization carried by these samples should be a thermoremanent magnetization. It should be also noted that the size of the magnetic grains observed is often too small.

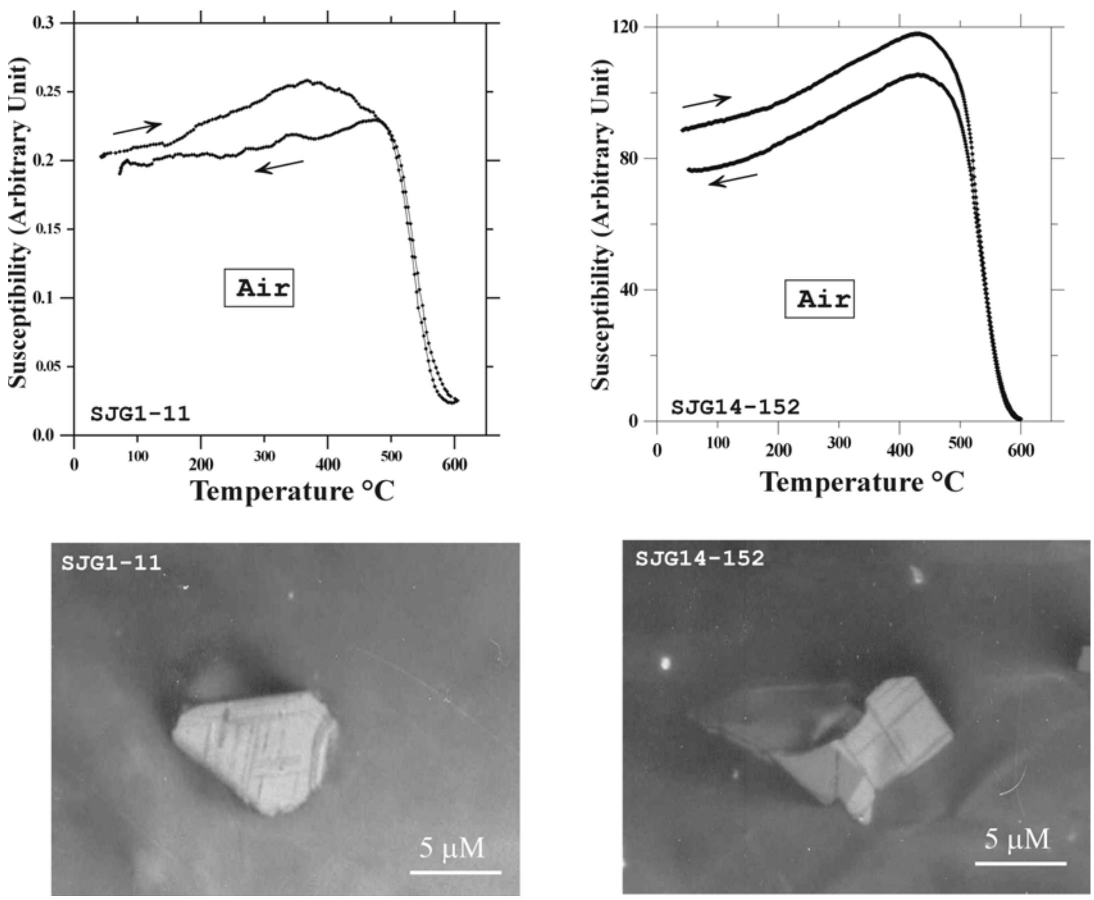

Figure 4. Top: Susceptibility versus temperature curves of representative samples selected for Thellier paleointensity experimenst. The arrows indicate the direction of heating and cooling curves. Bottom: Representative reflected light microphotograph of studied basalts, oil immersion, crossed nicols. Examples of ilmenite lamellae of the 'trellis' type in titanomagnetite transformed into Ti-poor titanomagnetite.

4.4. Hysteresis Experiments

Hysteresis measurements at room temperature (using AGFM-Micromag apparatus) were made on small chip specimens. These experiments were carried out in fields of up to 1.2 Tesla (Figure 5).

Figure 5. Typical examples of hysteresis loops (uncorrected) and associated isothermal remanence (IRM) acquisition curves of small chip samples from the studied volcanic units.

The hysteresis parameters (saturation remanent magnetization Jrs, saturation magnetization Js, and coercive force Hc) were calculated after correction for the paramagnetic contribution. Coercivity of remanence (Hcr) was determined by applying progressively increasing back-field after saturation. IRM (isothermal remanent magnetization) curves show (Figure 5) that saturation is reached in moderate fields (200 - 250mT), which points to some spinel phases (probably Ti-poor titanomagnetites) as predominant magnetic minerals. Judging from the ratios of hysteresis parameters (J_{rs}/J_s ranges from 0.22 to 0.42 and H_{cr}/H_c varies between 1.41 and 2.6), it seems that the magnetic particles fall in the small pseudo-single-domain grain size [Figure 6, *Day et al.,* 1977]. This probably indicates a mixture of multidomain and a significant amount of single-domain (SD) grains. However, we believe that the magnetic domain state estimation using room temperature hysteresis parameters in terms of the plot of magnetization ratio vs coercivity ratio has no resolution for most of natural rocks [*Goguitchaichvili et al.,* 2001c]. Thus, these data were not used to select the most promising samples for Thellier paleointensity experiments.

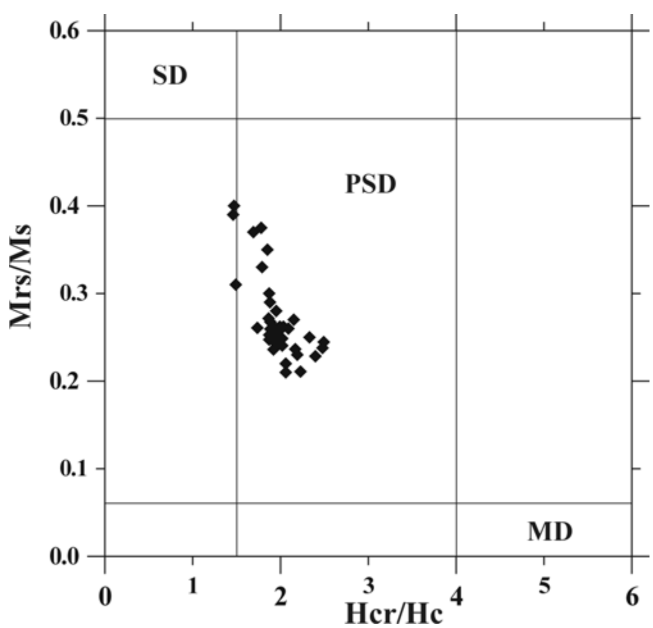

Figure 6. Day plot [*Day et al*., 1977] with the hysteresis parameters ratios.

In total only 73 samples providing the above described magnetic characteristics (i.e. yielding viscosity index lower than 5%, stable, univectorial remanent magnetization and close to reversible χ(T) curve) were selected for the paleointensity experiments.

5. Paleointensity Determination

While procedures for determining the direction of the paleomagnetic field are now more or less standardized, there still exist significant inter-laboratory differences on how to best obtain reliable estimates of the paleomagnetic field intensity [*Valet and Herrero-Bervera*, 2000]. The Thellier and Thellier technique and all derived approaches, specially *Coe et al*'s [1978] modification, received the largest audience last years [*Juarez et al*., 1998, *Selkin and Tauxe,* 2000, *Alva-Valdivia et al*., 2001, *Morales et al.*, 2001]. This technique was used in present study to determine absolute geomagnetic paleointensity. The method involves heating samples twice at each temperature step – once in zero magnetic field to remove a portion of NRM and once in a controlled field to determine the partial thermoremanence (pTRM) gained. The ratio of NRM lost to pTRM gained is proportional to the ancient field. The heatings and coolings were made in low pressure (better than 10^{-2} mbar) using MDT80 paleointensity oven and the laboratory field set to 30 µT. Ten temperature steps were distributed between room temperature and 560°C. Temperature reproducibility between two steps was in general better than 2°. The pTRM/NRM checks were performed after every second step throughout the experiments (Figure 7a).

Figure 7a. The representative NRM-TRM plots (so-called Arai-Nagata plots) and associated orthogonal vector diagrams from Baja California samples. In the orthogonal diagrams (in sample coordinates) we used same notations as in the figure 3.

The acceptance criteria for individual paleointensity determinations we used are basically similar to those reported in *Goguitchaichvili et al*., [1999a, 2000b, 2001b, 2002] and *Riisager et al*., [1999, 2000, 2002] and can be described as follow:

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

1. Obtained from at least 6 NRM-TRM points corresponding to a NRM fraction larger than about 1/3 [*Coe's* 1978 quality factor f >0.35]. This criteria are more restrictive

than those used by *Teanby et al*., [2002]; *Tanaka and Kono*, [2002] and *Laj et al*., [2002].

2. Yielding quality factor q [*Coe et al*., 1978] of about 4.5 or more. In very few cases (samples SI3_49A, SI3_36, SJG14_146A, SJG7_76B), we accepted determinations with slightly lower q factors because the intensity was found close to the site-mean value and f was above 0.45 (Table 2).
3. Positive pTRM checks (at least two) - We define pTRM checks as positive if the repeated pTRM value agrees with the first measurement within 10%. Because the small (from room temperature up to 250 °C) pTRMs are hard to measure precisely on the background of the full NRM/TRM, we were forced to allow some larger deviation of pTRM checks (within 15%).
4. 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 towards the direction of applied laboratory field was observed. To better illustrate this point, we calculated the ratio of potential CRM(T) to the magnitude of NRM(T) for each double heating step in the direction of the laboratory field during heating at T [*Goguitchaichvili et al*., 1999c]. The values of angle γ [the angle between the direction on characteristic remanent magnetization (ChRM) obtained during the demagnetization in zero field and that of composite magnetization (equal to NRM(T) if CRM(T) is zero) obtained from the orthogonal plots derived from the Thellier paleointensity experiments] are reported in table 2. For accepted determinations γ values are all < 10° which attest that no significant CRM was acquitted during the laboratory heatings

[*Goguitchaichvili et al.*, 1999c]. This approach is probably more restrictive than simple calculation of angle between the characteristic directions determined by Thellier experiments and by thermal demagnetization in zero field.

Only 33 samples, from 10 individual lava flows, yielded acceptable paleointensity estimates (Figure 7a, b and c).

Figure 7b. (continued)

Figure 7c. (continued)

For these samples the NRM fraction *f* used for determination ranges between 0.37 to 0.81 and the quality factor *q* varies from 24.4 to 4.5 (generally more than 5). The obtained paleointensity values are more or less reasonably coherent within a flow, generally varying by 10-20% from each mean.

The principal reason for rejecting a paleointensity determination was a typical ‘concave-up’ behavior [*Levi et al.,* 1977] detected in some cases (Figure 7d).

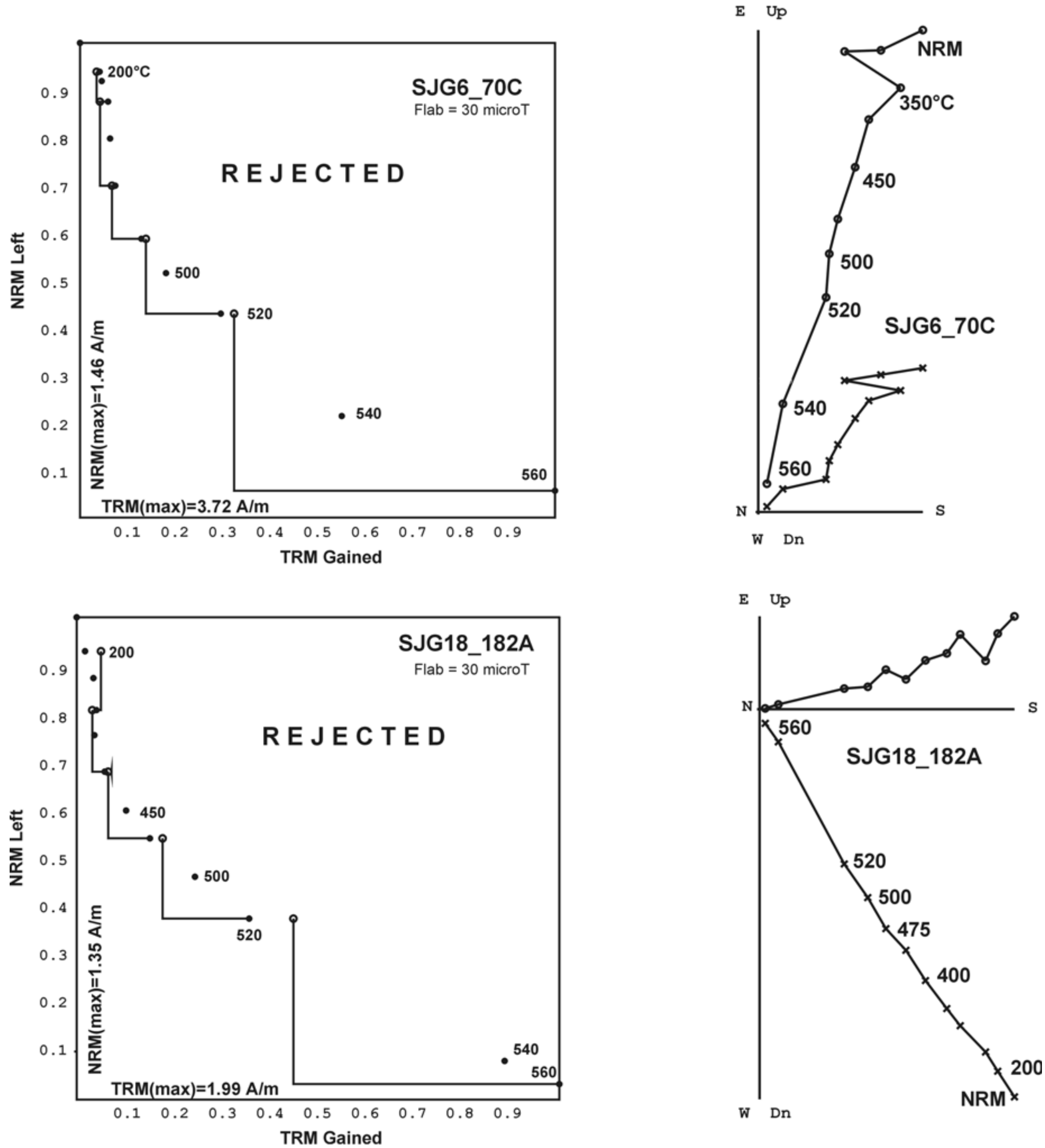

Figure 7d. (continued)

Up to 400° an important loss of NRM without any noticeable TRM acquisition but with positive pTRM checks is observed in Figure 7d. This phenomenon can be due to irreversible variations of coercive force [*Kosterov and Prévot,* 1998] at low temperature and can be interpreted as transformation from a single-domain 'metastable' state to multidomain state which results in a large NRM loss without any correlated TRM acquisition during the subsequent cooling. In our opinion, this point is a more complex puzzle in paleointensity whichis still not fully understood. In any case, these samples were not included in subsequent analyses.

Table 2. Paleodirectional and Paleointensity results

Site	*Dec(°)*	*Inc(°)*	*a_{95}(°)*	*k*	*Sample*	*N*	*T_{min}-T_{max}* °C	*γ(°)*	*f*	*g*	*q*	*$F_E \pm \sigma(F_E)$* (μT)	*VDM* $10^{22}Am^2$	*VDMe* $10^{22}Am^2$
SI3	349.2	27.2	2.9	135	SI3_36	7	350-540	2.3	0.45	0.78	4.9	34.1 ± 3.2	8.1	***7.3 ± 0.7***
					SI3_49A	6	400-540	2.5	0.45	0.78	4.5	30.5 ± 2.1	7.24	
					SI3_50	7	350-540	5.3	0.76	0.79	10.9	28.2 ± 1.5	6.67	
SJG1	1.9	42.6	3.2	98	SJG1_11	9	250-560	3.5	0.79	0.80	11.7	17.4 ± 1.0	3.65	***3.8 ± 0.6***
					SJG1_17	7	350-540	4.5	0.61	0.78	7.6	21.4 ± 1.6	4.48	
					SJG1_19	6	400-540	9.4	0.53	0.73	5.2	15.4 ± 1.1	3.23	
SJG4	335.1	50.4	6.7	60	SJG4_42A	9	200-500	2.1	0.66	0.81	12.9	50.5 ± 2.2	9.73	***8.8 ± 0.8***
					SJG4_47A	7	300-520	1.8	0.58	0.72	6.5	42.8 ± 2.4	8.24	
					SJG4_41A	7	250-500	3.9	0.53	0.73	5.7	43.6 ± 2.6	8.4	
SJG6	47.4	44.4	3.5	183.4	SJG6_74A	7	350-540	8.8	0.62	0.81	9.9	39.2 ± 2.0	8.06	
SJG7	349	55.2	4.1	104	SJG7_78B	7	350-540	3.6	0.69	0.79	9.1	38.5 ± 2.4	6.97	***7.7 ± 1.0***
					SJG7_76B	7	350-540	7.3	0.47	0.78	4.8	46.4 ± 4.6	8.4	
SJG12	11.2	41.6	2.7	201	SJG12_120B	7	350-540	4.1	0.65	0.71	9	23.4 ± 1.2	4.95	***5.8 ± 1.1***
					SJG12_117B	7	350-540	6.6	0.71	0.69	5.3	31.1 ± 2.7	6.58	
SJG14	337.2	48.1	3.2	166	SJG14_138B	8	300-540	3.1	0.65	0.77	7.2	37.5 ± 2.7	7.42	***6.8 ± 1.0***
					SJG14_141A	7	350-540	1.9	0.63	0.78	8.9	30.8 ± 2.2	6.1	
					SJG14_142A	7	400-560	1.5	0.78	0.76	12.4	34.3 ± 1.7	6.79	
					SJG14_143A	8	350-560	3.6	0.81	0.77	19.4	34.7 ± 1.1	6.87	
					SJG14_144A	8	350-560	4.1	0.74	0.76	10.7	36.3 ± 1.8	7.18	
					SJG14_145A	8	350-560	2.1	0.72	0.8	9.1	37.8 ± 2.4	7.48	
					SJG14_146A	7	350-540	4.6	0.46	0.81	4.7	40.7 ± 3.2	8.08	
					SJG14_152C	7	350-540	4.9	0.44	0.79	5.7	23.8 ± 1.9	4.71	
SJG15	349.7	48.6	2.5	294	SJG15_154B	6	350-520	5.5	0.37	0.71	6.2	29.2 ± 2.0	5.74	***7.0 ± 1.2***
					SJG15_157A	7	350-540	2.6	0.57	0.82	7.2	38.3 ± 3.1	7.53	
					SJG15_158B	7	350-540	2.5	0.57	0.72	7.1	32.6 ± 1.8	6.41	
					SJG15_161E	6	350-520	3.3	0.47	0.81	5.6	42.6 ± 3.2	8.38	
SJG16	349.4	44.6	2.2	385	SJG16_168B	7	350-540	4.1	0.48	0.77	6.2	34.1 ± 1.8	7.65	***7.4 ± 0.3***
					SJG16_166C	6	350-520	7.6	0.37	0.74	7.8	32.1 ± 1.1	7.2	
SJG19	345.8	58.1	1.6	363	SJG19_189A	7	350-540	2.2	0.48	0.77	22.5	30.1 ± 0.4	5.28	***5.8 ± 0.5***
					SJG19_193A	7	350-540	2.0	0.41	0.78	24.4	29.9 ± 0.4	5.25	
					SJG19_194A	6	350-520	3.6	0.42	0.77	9.2	35.5 ± 1.3	6.23	
					SJG19_195A	6	350-520	5.8	0.43	0.76	5.6	36.6 ± 2.1	6.42	
					SJG19_198A	6	350-520	3.2	0.39	0.76	9	32.4 ± 1.1	5.65	

Paleodirectional and Paleointensity resuls from Baja California Miocene volcanic rocks:**Dec**: Declination, **Inc**: Inclination, **k** and **α_{95}**: Precision parameter and radius of confidence cone, **N** is the number of NRM-TRM points used for paleointensity determination, **Tmin-Tmax** is the temperature interval used, γ is the angle between the direction on characteristic remanent magnetization (ChRM) obtained during the demagnetization in zero field and that of composite magnetization (equal to NRM(T) if CRM(T) is zero (see text and *Goguitchaichvili et al.*, 1999c), **f**, **g** and **q** are the fraction of extrapolated NRM used, the gap factor and quality factor (*Coe et al.*, 1978) respectively. **Fe** is the individual paleointensity estimate with standard deviation, **VDM** and **VDMe** are individual and average virtual dipole moments.

6. Discussion and Main Results

The mean paleointensity values per flow obtained in this study range from 18.1 ± 3.1 to 45.6 ± 4.3 μT and the corresponding Virtual Dipole Moments (VDMs) range from 3.8 ± 0.6 to 8.8 ± 0.8 (10^{22} Am^2) with a mean value of 6.6 ± 1.7 x 10^{22} Am^2, which is about 85% of the present geomagnetic axial dipole 7.8 x 10^{22} Am^2 [after *Barton et al*., 1996]. This value is very similar to those recently reported from Central Mexico late Miocene volcanics [*Goguitchaichvili et al.,* 2000a] and Steens Mountain basalts (Table 3). However, The absolute paleointensities obtained from Central France [*Riisager et al*.,, 2000] and Oceanic Basalts [*Juarez et al.*, 1998] are significantly lower.

Two (of three) determinations obtained by *Riisager et al*., [2000] are made on the baked sediments. It can not be ascertained that the remanence carried by these units is of thermoremanent origin, because we still don't have any strict magnetic criteria to distinguish between TRM and CRM (chemical remanent magnetization). Although, *McClelland* [1996] theoretically predicted that CRM should yield a concave up Arai diagrams, *Körner et al*., [1998] and *Goguitchaichvili et al*., [1999b] do not observe this concavity.

Because, it was recently shown that the submarine basaltic glass probably contains a grain growth CRM [*Heller et al*., 2002], we calculated the mean VDM for late Miocene by two different ways: 1) considering SBG results (Table 3) and 2) discarding them. In both cases, however quite similar results are obtained. The average VDM value for late Miocene is 6.1 ± 1.5 (10^{22} Am^2), which is higher than average paleofield for the period 5-160 Ma [4.2 ± 2.3 x 10^{22} Am^2 after *Juarez and Tauxe*, 2000]. For Oligocene time (applying the same selection criteria to paleointensity results as in present study) we obtained mean VDM of 4.6 ± 0.7 x 10^{22} Am^2 (Table 3), which is comparable to values found by *Juarez and Tauxe*, [2000]. Reliable paleointensity data are relatively abundant for Plio-Pleistocene time (Table 3 and Figure 8). They yield an average dipole moment of 6.8 ± 1.1 x 10^{22} Am^2, which is higher comparing to mean VDM for the period spanning 0.3-5 Ma [5.5 ± 2.4 x 10^{22} Am^2 after *Juarez and Tauxe*, 2000]. Given the large dispersion and the very poor distribution of reliable absolute intensity data (Figure 8), it is hard to draw any firm conclusions regarding the time evolution of the geomagnetic field. As showed by *Guyodo and Valet* [1999] a very large number of records of absolute intensity, reinforced by numerous and precise ages, are necessary to provide a reasonable picture of the field variations.

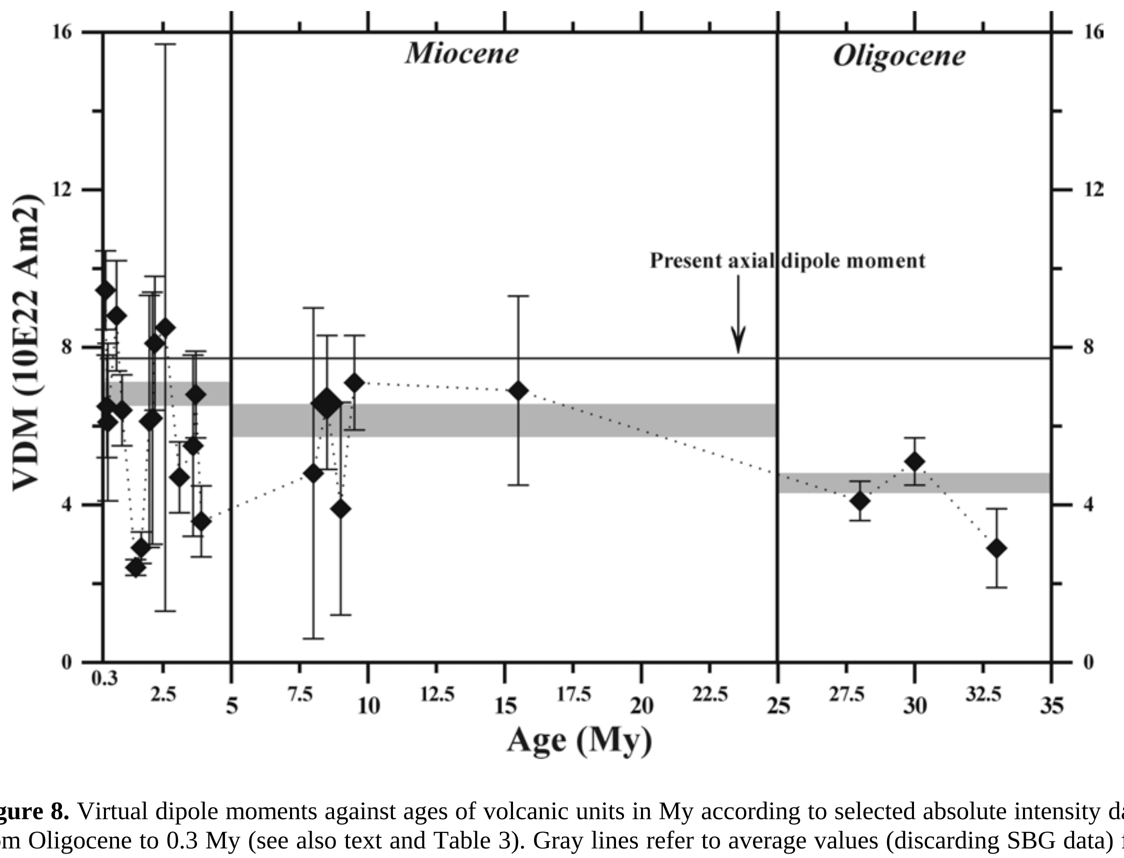

Figure 8. Virtual dipole moments against ages of volcanic units in My according to selected absolute intensity data from Oligocene to 0.3 My (see also text and Table 3). Gray lines refer to average values (discarding SBG data) for each geologic period as shown in Table 3. Squares denote to the mean value of several consecutive lava flows (This study – big symbol). For Plio-Pleistocene time, however some points represent the results from single cooling units see [*Alva-Valdivia et al.*, 2001].

Table 3. Average virtual dipole moments

Plio-Pleistocene

Region	***Age***	***N***	***VDM***	***S.D.***	***Reference***
	My		10^{22} Am^2		
Oceanic Basalts	3.9 - 2	3	**5.2**	1.8	*Juarez & Tauxe*, 2000
Oceanic Basalts	3.9 – 0.4	3	**4.9**	2.8	*Selkin & Tauxe*, 2000
East Eifel	0.49 – 0.35	7	**6.5**	1.3	*Schnepp*, 1996
Georgia	3.8 – 3.6	3	**6.8**	1.1	*Goguitchaichvili et al*. 2000a
Mexico	2.2 – 0.8	4	**8.0**	1.1	*Alva-Valdivia et al.* 2001
SW Iceland	~ 2.58	3	**8.5**	7.2	*Tanaka et al*. 1995
SW Iceland	~ 2.11	14	**6.2**	2.4	*Goguitchaichvili et al*. 1999a
West Eifel	0.55 – 0.4	24	**6.1**	2.0	*Schnepp & Hradetzky*, 1994
Georgia	~ 3.6	6	**5.5**	2.3	*Goguitchaichvili et al.* 2000b

Total ***6.4*** ***1.2***
Total* ***6.8*** ***1.1***

Miocene

Region	***Age***	***N***	***VDM***	***S.D.***	***Reference***
	My		10^{22} Am^2		
Baja California	6.0 - 11.7	6	**6.6**	1.7	This Study
Central Mexico	8.1 - 10.3	4	**7.1**	1.2	*Goguitchaichvili et al.* 2000c
Central France	8.6 - 9.7	3	**3.9**	2.7	*Riisager et al.*1999
Oceanic Basalts	6.4 - 10.3	2	**4.8**	4.2	*Juarez et al.* 1998
North America	~ 15.5	10	**6.9**	2.4	*Prévot et al.* 1985

Total ***5.9*** ***1.4***
Total* ***6.1*** ***1.5***

Oligocene

Region	***Age***	***N***	***VDM***	***S.D.***	***Reference***
	My		10^{22} Am^2		
Ethiopia	~ 30	2	**5.1**	0.6	*Riisager et al.* 1999
Oceanic Basalts	32 - 35	5	**2.9**	1.0	*Juarez et al.* 1998
North. Mexico	27 - 29	3	**4.1**	0.5	*Goguitchaichvili et al.* 2001a

Total ***4.0*** ***1.1***
Total* ***4.6*** ***0.7***

Average virtual dipole moment (10^{22} Am^2) for various regions since Oligocene (from selected Thellier paleointensity data, see text) to 0.3 My. N refers to the number of cooling units used to calculate the average VDM. Asterisks refer to the mean values calculated discarding SBG results.

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Absolute Paleointensity of the Earth's Magnetic Field During Jurassic: Case Study of La Negra Formation (Northern Chile)

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Abstract

We carried out a detailed rock-magnetic and paleointensity study of the ~ 187 Ma volcanic succession from northern Chile. A total of 32 consecutive lava flows (about 280 oriented standard paleomagnetic cores) were collected at the Tocopilla locality. Only twenty-six samples with apparently preserved primary magnetic mineralogy and without secondary magnetization components were pre-selected for Thellier paleointensity determination. Eleven samples coming from 4 lava flows yielded reliable paleointensity estimates. The flow-mean Virtual Dipole Moments range from 3.7 ± 0.9 to 7.1 ± 0.5 (10^{22} Am^2). This corresponds to a mean value of $5.0 \pm 1.8 \times 10^{22}$ Am^2, which is in reasonably good agreement with other comparable quality paleointensity determinations from the middle Jurassic. Given the large dispersion and the very poor distribution of reliable absolute intensity data, it is hard to draw any firm conclusions regarding the time evolution of the geomagnetic field.

Key words: Paleointensity, Paleomagnetism, Jurassic, northern Chile.

Introduction

Geomagnetic paleointensity data are critical for understanding the workings of the Earth's dynamo. Coe et al. (2000) recently suggested that absolute intensity should be a fundamental constraint in numerical models that promise to provide unprecedented insight into the operation of the geodynamo. However, absolute paleointensity data are still scarce and cannot be yet used to document the general characteristics of the Earth's magnetic field (Selkin and Tauxe, 2000). Reliable paleointensity values are generally much more difficult to obtain than reliable directional data, in part, because only volcanic rocks which satisfy some very specific magnetic criteria (Kosterov and Prévot, 1998, Calvo et al., 2002) can be used for absolute paleointensity determination.

Judging from the existing paleointensity database (ftp://ftp.dstu.univ-montp2.fr/pub/paleointb, see also Perrin et al., 1998) only large-scale features can be resolved, such as the pronounced lowering of the Earth's Virtual Dipole Moment (VDM) in the whole Mesozoic and part of Paleozoic (Prévot et al., 1990), during which the dipole structure of the field was preserved (Perrin and Shcherbakov, 1997). However, high geomagnetic intensities have been recently reported (Tarduno et al., 2001; Goguitchaichvili et al., 2002) for the middle Cretaceous and thus the reliability of Mesozoic Dipole Low (MDL) may be questioned. In order to better constraint the Mesozoic geomagnetic field strength, we carried out Thellier paleointensity experiments on the ~ 187 Ma old La Negra volcanic succession in northern Chile. Obtaining new paleointensity and paleodirectional data from Southern America volcanics is critical because the geographic distribution of paleointensity data is still very uneven (e.g. Tanaka et al., 1995). In particular very few data are available from the Southern hemisphere, which impedes an accurate analysis of the fine-scale changes in the statistical characteristics of paleosecular variations (Jacobs, 1994).

Sampling Details

The La Negra Formation (Figure 1a and b) comprises a thick pile of lava flows, largely of high-K basaltic andesites, with thin intercalated sediments. It seems that they erupted in an extensional environment (Rogers et al., 1989) and that, together with their chemical similarity to the Puente Piedra volcanics in Peru (Atherton et al., 1983) and the lavas at Bustamante Hill in Central Chile (Levi et al., 1982), suggests they were erupted in an ensialic back-arc basin rather than in an actual volcanic arc.

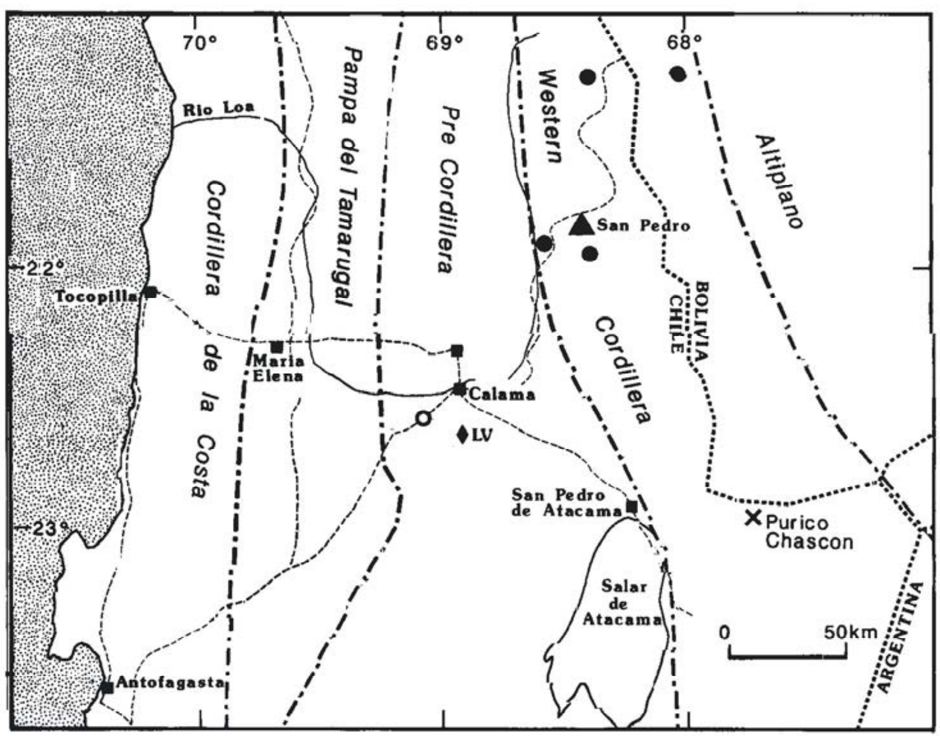

Figure 1. A) Sketch map of northern Chile illustrating the locality of paleomagnetic sampling (modified from Rogers et al., 1989). 1 – La Negra Formation, 2 and 3 are Gatico (~ 158 Ma) and Tocopilla (~ 155 Ma) intrusions respectively

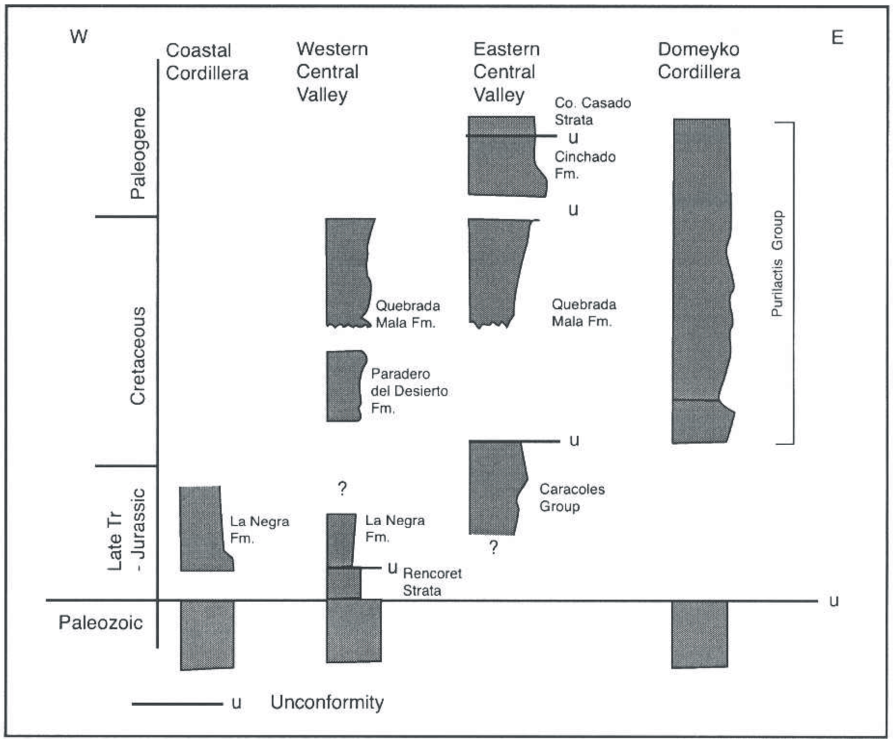

Figure 1. B) Simplified stratigraphy of Antofagasta region, northern Chile, with the principal names of geological formations (redrawn from Arriagada et al., 2002).

Paleomagnetic sampling was done near the town of Tocopilla (Figure 1). Rogers et al. (1989) proposed a mean age of 187 Ma as the best estimate of the time of emplacement of the lower La Negra volcanic sequence. It is only regrettable that these authors do not report the error on their K-Ar age. We note however, that the sample dated by Rogers et al., (1989) most probably, belongs to site NE06 of our collection (lower part of La Negra sequence, Alva-Valdivia et al., 2003). In total, about 280 oriented samples belonging to 32 consecutive lava flows were collected. Commonly, the outcrops extend laterally over a few tens of meters and in these cases we drilled typically 6-11 standard paleomagnetic cores per flow. The samples were distributed throughout each flow both horizontally and vertically in order to minimize effects of block tilting and lightning. Cores were obtained with a gasoline-powered portable drill, and then oriented with both magnetic and sun compasses.

Rock-Magnetic Characteristics of Selected Samples

Magnetic characteristics of typical samples selected for Thellier paleointensity measurements are summarized in Figure 2 (upper part) and could be described as follows:

1. Selected samples carry essentially a single and stable component of magnetization, observed upon alternating field (Figure 2a, sample NE2-011C) treatment. A generally minor secondary component, probably of viscous origin, was present but was easily removed. The median destructive fields (MDF) range mostly in the 40-60 mT interval, suggesting the existence of small pseudo-single to single magnetic domain grains as remanence carriers, which are suitable material for the Thellier paleointensity study (Dunlop and Ozdemir, 1997).

2. Continuous low-field susceptibility measurements with temperature show the presence of a single ferrimagnetic phase with Curie point compatible with Ti-poor titanomagnetite. However, the cooling and heating curves are sometimes not perfectly reversible (Figure 2b, sample NE2-011D) probably due to the heatings in air. In addition, hysteresis measurements were carried out on all samples (not shown). However, we believe that the magnetic domain state estimation using room temperature hysteresis parameters in terms of the plot of magnetization ratio vs coercivity ratio has no resolution for most of natural rocks (Goguitchaichvili et al., 2001a). Thus, these data were not used to select the most promising samples for Thellier paleointensity experiments.

In all we selected 26 samples for the paleointensity experiments having the above-described magnetic characteristics. Remaining samples may be divided in two groups (Figure 2, middle and lower parts) after their magnetic properties: i) the samples yield reasonably reversible continuous susceptibility curves (Figure 2b, sample NE24-168A) associated with the remanence dominated by strong secondary magnetization (Figure 2a, sample NE24-171A; ii) The k-T curves of these samples yield the evidence of two ferrimagnetic phases during heating (Figure 2c, sample NE9-055). The lower Curie point ranges between 350-400°C, and the highest one is about 580°C. The cooling curve shows only a single phase, with a Curie temperature close to that of magnetite. Such irreversible k-T curves can be explained by titanomaghemite, which probably transformed into magnetite (Readman and O'Reilly, 1972; Özdemir, 1987) during heating. It is possible that these samples carry chemical remanent magnetization. In addition, this behavior is accompanied by very unstable multivectorial magnetization (new version of Figure 2a, sample NE09-055B). It is obvious that these samples cannot be used to determine absolute geomagnetic paleointensity.

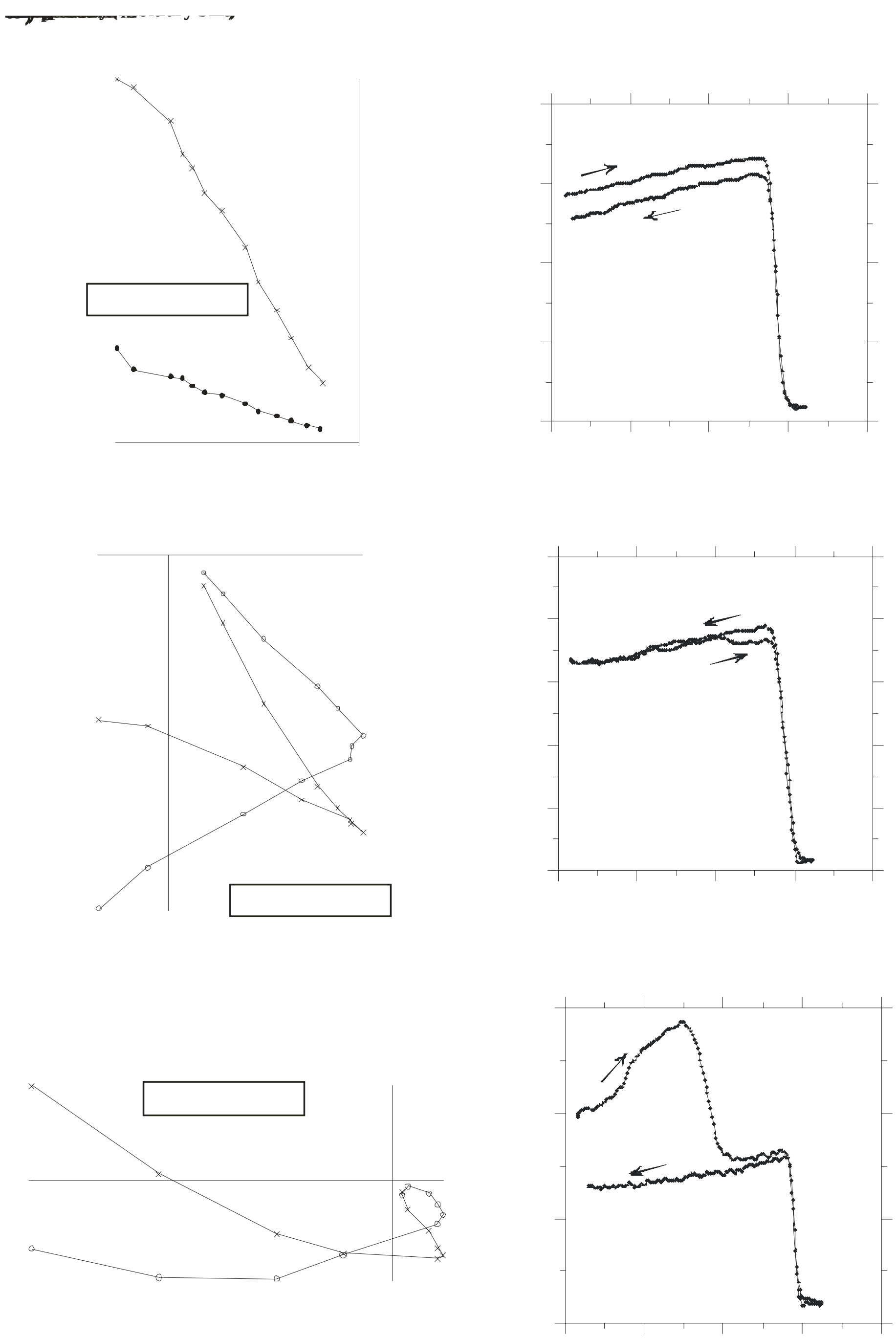

Figure 2. Summary of the magnetic characteristics of the typical samples, selected (class 1) and unselected (class 2 and 3 see also text) for Thellier paleointensity experiments. A) Orthogonal vector plots of stepwise alternating field demagnetization (stratigraphic coordinates). The numbers refer to peak alternating fields in mT. ● – projections into the horizontal plane, x – projections into the vertical plane. B) Susceptibility versus temperature curves. The arrows indicate the heating and cooling curves.

As part of their efforts to study the details of the volcanic stratigraphy and paleotectonics of Antofagasta region (northern Chile), Alva-Valdivia et al., (2003) recently carried out a detailed paleomagnetic study of La Negra Formation. The reliable paleodirections were obtained for 25 sites among 32 sampled (Figure 3). These directions are considered to be of primary origin because of the occurrence of almost antipodal normal and reversed polarities (Figure 3). In addition, unblocking temperature spectra and relatively high coercivities point to 'small' pseudo-single domain magnetic structure grains as responsible for remanent magnetization. The secondary magnetizations were successfully removed in most cases using the alternating field demagnetization technique.

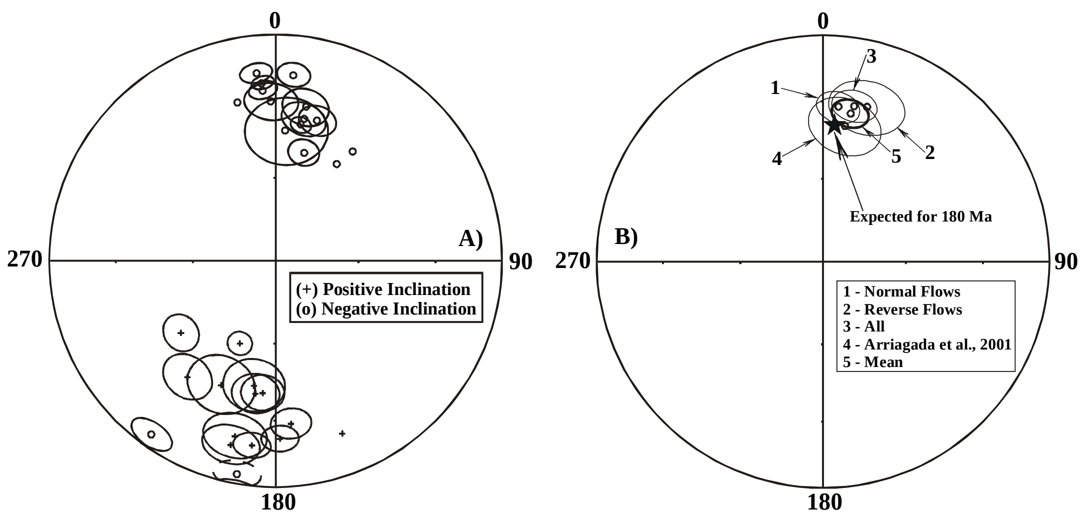

Figure 3. A) Equal area projections of the flow mean characteristic paleodirections for La Negra Formation. Circles/Crosses denote to the negative/positive inclination. B) Equal area projection of the mean directions for Antofagasta-Tocopilla volcanic units (Alva-Valdivia et al., 2003), expected value from Roperch and Carlier, 1992).

The mean paleodirection obtained in this study is I=-30.7, D=11.3°, k=18, α_{95}=6.8°, N=25 (Figure 3). The previous paleomagnetic study provide basically similar directions with I=-39.2°, D=9.1°, k=11, α_{95}==12°, N=15 (Arriagada et al., 2002). Combining both data we obtained a well-defined mean paleomagnetic direction with I=-33.9°, D=10.5°, k=15, α_{95}=6.1°, N=40. These directions are in good agreement with the expected paleodirections at about 180 Ma (Ropperch and Carlier, 1992, Figure 3b). This indicate that studied units do not show evidence for significant tectonic rotation. Fifteen sites yield reverse polarity magnetization and 11 flows are normally magnetized. The tentative direct magnetostratigraphic correlation suggests that La Negra volcanics have been emplaced during a relatively large time span of about 5 My (Alva-Valdivia et al., 2003).

Paleointensity Determination

Paleointensity experiments were performed using the Thellier method in its classic form (Thellier and Thellier, 1959). All heatings were made in a vacuum better than 10^{-2} mbar. All remanences were measured using both JR5A and JR6 spinner magnetometers. Paleointensity data are reported on the Arai-Nagata (Nagata at al., 1963) plot on Figure 4 and results are given in Table 1.

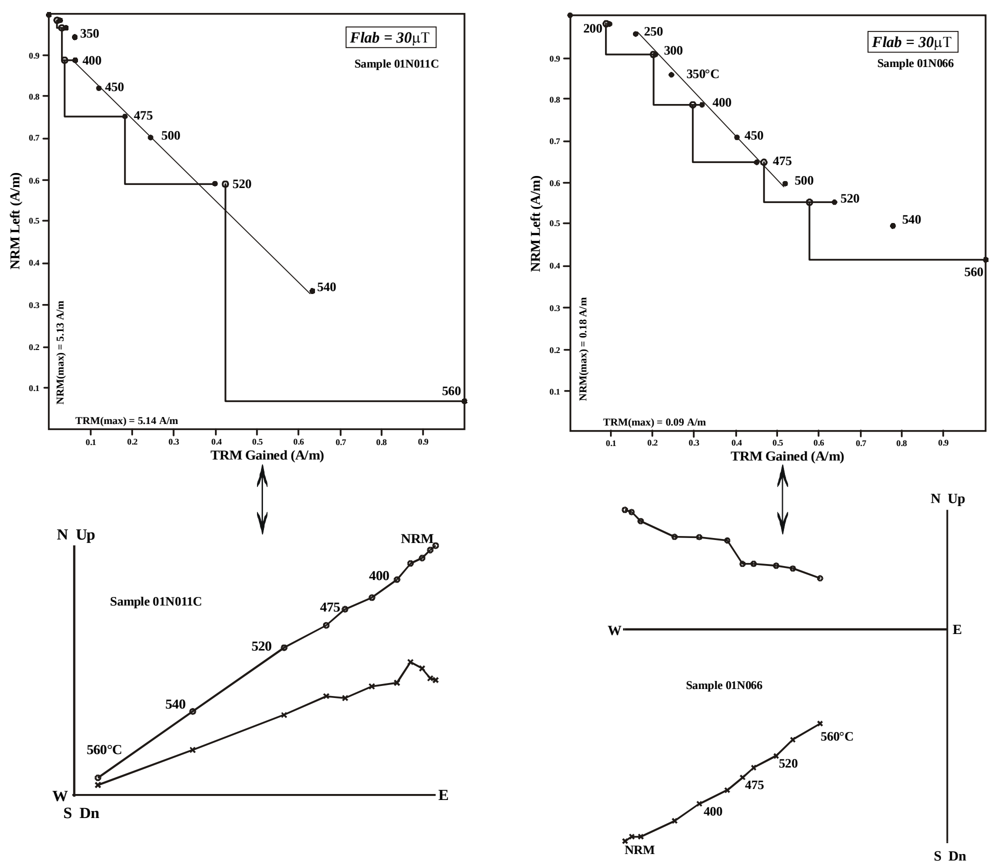

Figure 4. The representative NRM-TRM plots and associated orthogonal diagrams for La Negra samples. In the orthogonal diagrams we used same notations as in the figure 2a

Table 1

Paleodirections						Paleointensity										
Site	***n/N***	***Dec***	***Inc***	α_{95}	***k***	***Specimen***	***Inc'***	***Tmin-Tmax***	***N'***	***f***	***g***	***q***	γ	$F_E \pm \sigma(F_E)$	***VDM***	***VDMe***
		(°)	(°)	(°)			(°)									
NE02	6/6	14.8	-49.2	5.5	148	01N007C	-48.4	350-520	6	0.54	0.78	5.9	3.3	18.2 ± 1.7	3.59	**4.2 ± 0.5**
						01N008B	-49.6	400-520	5	0.55	0.78	8.4	4.4	21.5 ± 1.3	4.18	
						01N011C	-56.2	400-540	6	0.58	0.81	9.8	1.5	26.1 ± 2.6	4.69	
						01N012C	-54.3	450-560	6	0.75	0.79	6.5	2.9	24.0 ± 2.1	4.41	
NE11	6/6	203.4	57.3	4.3	159	01N062C	51.4	250-500	7	0.42	0.82	4.5	10.6	35.1 ± 3.1	6.68	**7.1 ± 0.5**
						01N063B	54.3	250-500	7	0.45	0.85	6.6	6.3	41.7 ± 3.0	7.67	
						01N066C	60.5	250-520	8	0.51	0.87	10.7	2.1	40.4 ± 1.5	6.87	
NE19	7/7	11.3	-36.2	6.9	76	01N142B	-38.6	300-520	7	0.45	0.86	9.4	0.8	13.8 ± 0.7	3.0	**3.7 ± 0.9**
						01N143C	-39.3	300-500	6	0.30	0.83	4.4	8.8	21.6 ± 1.7	4.67	
						01N145B	-44.2	300-500	6	0.35	0.87	6.3	5.3	16.3 ± 1.1	3.36	
NE20	6/6	5.4	-18.8	5.4	128	01N146B	-19.6	250-500	7	0.31	0.82	5.5	7.7	14.5 ± 0.7	3.59	

Table 1. Paleodirectional and Paleointensity results from La Negra volcanic lava flows: **N**, number of treated samples; **n**, number of specimens used for calculation, **Dec**: Declination, **Inc**: Inclination, **k** and α_{95}: Precision parameter and radius of confidence cone, **Inc'** is the individual inclination derived from Thellier paleointensity experiments, **N'** is number of NRM-TRM points used for paleointensity determination, **Tmin-Tmax** is the temperature interval used, **f, g** and **q** are the fraction of extrapolated NRM used, the gap factor and quality factor (Coe et al., 1978) respectively. γ is the angle between the direction on characteristic remanent magnetization (ChRM) obtained during the demagnetization in zero field and that of composite magnetization (equal to NRM(T) if CRM(T) is zero (see text and Goguitchaichvili et al., 1999a), **Fe** is the individual paleointensity estimate with standard deviation, **VDM** and **VDMe** are individual and average virtual dipole moments.

We accepted only determinations: (1) obtained from at least 5 NRM-TRM points corresponding to a NRM fraction larger than about 1/3 (Table 1), (2) yielding quality factor (Coe, et al., 1978) above 4.5, (3) with positive 'pTRM' checks i.e. the deviation of 'pTRM' checks were less than 15% and (4) with reasonably linear Zijderveld diagrams obtained from the paleointensity experiments. No significant deviation of NRM remaining directions towards the direction of applied laboratory field was observed. To better illustrate this point, we calculated the ratio of potential CRM(T) to the magnitude of NRM(T) for each double heating step in the direction of the laboratory field during heating at T (Goguitchaichvili et al., 1999a). The values of angle γ [the angle between the direction on characteristic remanent magnetization (ChRM) obtained during the demagnetization in zero field and that of composite magnetization (equal to NRM(T) if CRM(T) is zero) obtained from the orthogonal plots derived from the Thellier paleointensity experiments] are reported in table 1. For accepted determinations γ values are all < 10.6° which attest that no significant CRM was acquitted during the laboratory heatings (Goguitchaichvili et al., 1999a). This approach is probably more restrictive than simple calculation of angle between the characteristic directions determined by Thellier experiments and by thermal demagnetization in zero field.

Main Results and Discussion

Only 11 samples, coming from 4 individual lava flows, yielded acceptable paleointensity estimates. The almost 95 percent failure rate that we find in our study is not exceptional for a Thellier paleointensity study of old volcanic rocks, if correct pre-selection of suitable samples and strict analysis of the obtained data are made. The Thellier and Thellier (1959) method of geomagnetic absolute intensity determination, which is considered the most reliable one, imposes many restrictions on the choice of samples that can be used for a successful determination (Coe, 1967, Levi, 1977, Prévot et al., 1985, Pick and Tauxe, 1993a, Kosterov and Prévot, 1998, Calvo et al., 2002).

Although our results are not numerous, some credit should be given because of good technical quality determination, attested by the reasonably high Coe et al's (1978) quality factors. For the accepted determinations, the NRM fraction f used for determination ranges between 0.30 to 0.75 and the quality factor q varies from 4.4 to 10.7 (generally more than 5). The mean Virtual Dipole Moments (VDMs) are ranging from 3.7 ± 0.9 to 7.1 ± 0.5 (10^{22} Am^2). This corresponds to a mean value of $5.0 \pm 1.8 \times 10^{22}$ Am^2, which is in agreement with other comparable quality paleointensity determination for middle Jurassic (Kosterov et al., 1997; Selkin and Tauxe, 2000). The site NE20, which is representad only by single determination, was discarded to calculate mean VDM.

La Negra mean VDM is shown on Figure 5 together with 16 other selected mean VDMs and VADMs (virtual axial dipole moment) for the period 5-200 Ma.

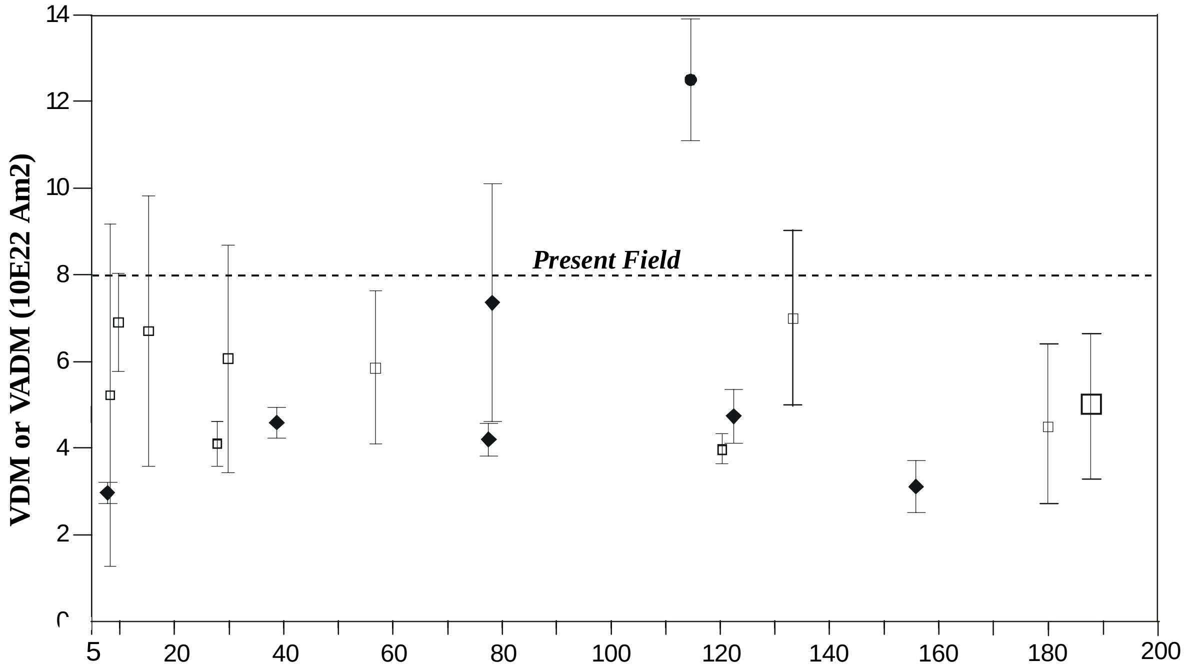

Figure 5. Evolution of mean virtual dipole moments (VDMs) and virtual axial dipole moments (VADMs) for 5 to 200 Ma. Open squares are mean VDM obtained from continental lava flows (Goguitchaichvili et al., 2000, 2001b, 2002; Prévot et al., 1985; Riisager et al., 1999, 2000, 2002; Zhu et al., 2001) (This study – big symbols). Diamonds are mean VADMs obtained from submarine basaltic glass (Selkin and Tauxe, 2000). Closed circle is mean VDM obtained single plagioclase crystals (Tarduno et al., 2001).

We selected data using quite modest criteria (Riisager et al., 2002) demanding a) the mean based on more than 9 successful determinations from at least three cooling units, b) no transitional data and c) paleointensity estimates obtained with Thellier method with pTRM checks. The perhaps most striking observation from inspecting Figure 5 is that, in spite of the many published paleointensity studies, there exists only 17 paleomagnetic dipole moment estimates in the 5-200 Ma period that fulfill the above mentioned modest criteria. Among them 6 are from submarine basaltic glass, 1 from single plagioclase crystal and remaining VDMs come from whole-rock samples. The excellent technical quality of paleointensity data obtained from basaltic glasses, first underlined by Tauxe and Pick (1993a,b) and later by Juarez et al. (1998 and 2000) is not by itself a proof of the geomagnetic validity of the paleostrength found. Goguitchaichvili et al. (1999b) showed that the magnetite found in this kind of material may not be of magmatic origin but crystallizes at rather low temperatures as a result of either glass demixtion or alteration. In such a case, the Thellier paleointensity results would underestimate the actual paleofield strength. More data are needed before the intriguing differences between different materials can be put into a geomagnetic context.

Given the large dispersion and the very poor distribution of reliable absolute intensity data, it is hard to draw any firm conclusions regarding the time evolution of the geomagnetic field. As showed by Guyodo and Valet (1999) a very large number of records of absolute intensity, reinforced by numerous and precise ages, are necessary to provide a reasonable picture of the field variations trough time.

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On the Reliability of Mesozoic Dipole Low: New Absolute Paleointensity Results from Paraná Flood Basalts (Brazil)

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Abstract

Thellier paleointensity experiments were carried out on Early Cretaceous Paraná Flood Basalts. Forty-two samples from 11 lava flows yielded apparently reliable absolute intensity determinations. The mean paleointensity values per flow obtained in this study are ranging from 19.4 ± 4.0 to 46.7 ± 7.0 μT and the corresponding Virtual Dipole Moments are ranging from 4.0 ± 0.6 to 10.5 ± 1.1 (10^{22} Am^2). These yield a mean value of 7.2 ± 2.3 x 10^{22} Am^2, which is about 92% of the present geomagnetic axial dipole (7.8 x 10^{22} Am^2 after Barton et al., 1996). Our results in conjunction with Tarduno et al's (2001) data suggest that the paleostrength during early Cretaceous period might be comparable or even much higher than recent field intensity and not 'anomalousely low' as previously suggested. Thus, the Mesozoic Dipole Low may be probably considered as unreliable.

Key Words: Paleointensity, Mesozoic, Paraná Flood Basalts, Brazil.

Introduction

Variation of paleointensity over geologic time may indicate modulation of geodynamo action in the core by the convective state of the lower mantle. Thus, determinations of the absolute intensity of the Earth's magnetic field in the past are decisive for understanding the processes in the core that give rise to the geomagnetic field, and how and why the Earth's magnetic field reverses polarity. Coe et al. (2000) recently suggested that absolute intensity should be a fundamental constraint in numerical models that promise to provide unprecedented insight into the operation of the geodynamo. Despite of about forty years of research, paleointensity data are scarce (Selkin and Tauxe, 2000) and they cannot be yet used to document a long-term variation in the intensity of the Earth's magnetic field through geological time with sufficient detail.

Prévot et al. (1990) based on paleointensity data compilation since the Triassic period (Bol'shakov and Solodovnikov, 1983) first underlined the existence of relatively low field during Mesozoic time (mainly from 180 to 120 Ma). The dipole strength was found only one third of the Cenozoic value, prevailing during most of Mesozoic times. Kosterov et al. (1998), Perrin et al. (1991) and Perrin and Shcherbakov (1997) found further evidence of this Mesozoic Dipole Low (MDL) by detailed experimental and statistical analyses of the paleointensity records. More recently, based on new high-technical-quality data from submarine basaltic glasses, Juarez et al. (1998) argued that the average dipole moment of the Earth over the past 160 Ma was only half of present day field, suggesting that MDL was not 'low' but of average intensity. Basically, the same conclusion was reached by Selkin and Tauxe (2000). High geomagnetic intensities of $(12.5 \pm 1.4) \times 10^{22}$ Am^2 have been reported (Tarduno et al., 2001) for the interval 113

– 116 Ma, which are consistent with some inferences from computer simulations (Glatzmaier et al., 1999). Zhu et al. (2001) found relatively low dipole-field intensity just prior to the Cretaceous normal superchron. Thus, more reliable paleointensity data are strongly needed to asses the validity of the MDL hypothesis.

In this paper, we present new paleointensity results for the Paraná Flood Basalts (southern Brazil) erupted between 133 and 132 Ma (Renne et al., 1992, 1993; Tamrat et al., 1999).

Ages and Sampling Details

The Paraná Magmatic Province (PMP) represents one of the world's largest volumes of Mesozoic continental flood basalt, covering an area about 1.2 x 10^6 km^2, located in southern Brazil (mainly), Uruguay, Paraguay and Argentina. The Paraná lavas (Serra Geral Formation) overlie the Botucatu formation (Jurassic-Cretaceous), which is composed of typical aeolian sandstones representing the top of the Gondwana sequence. Renne et al., (1992, 1996a,b) showed that the entire PMP erupted in a very narrow age interval (mainly 133-132 Ma) although slightly older ages have been reported (e.g. Turner et al. 1994).

The PMP is divided into three major parts by the Rio Uruguay and Rio Piquiri tectonic lineaments (Figure 1) which existed since Devonian. This division is not only tectonic but also can be traced by geochemical data (Stewart et al., 1996). Tholeiitic basalts are the dominant rock type, however, volcanic suites may also include some acid rocks (Ernesto et al., 1990, Tamrat et al., 1999). Paleomagnetic sampling was done in the central part of PMP (Figure 1) between both major lineaments. Cores were drilled in the field and oriented in most cases with both magnetic and sun compasses.

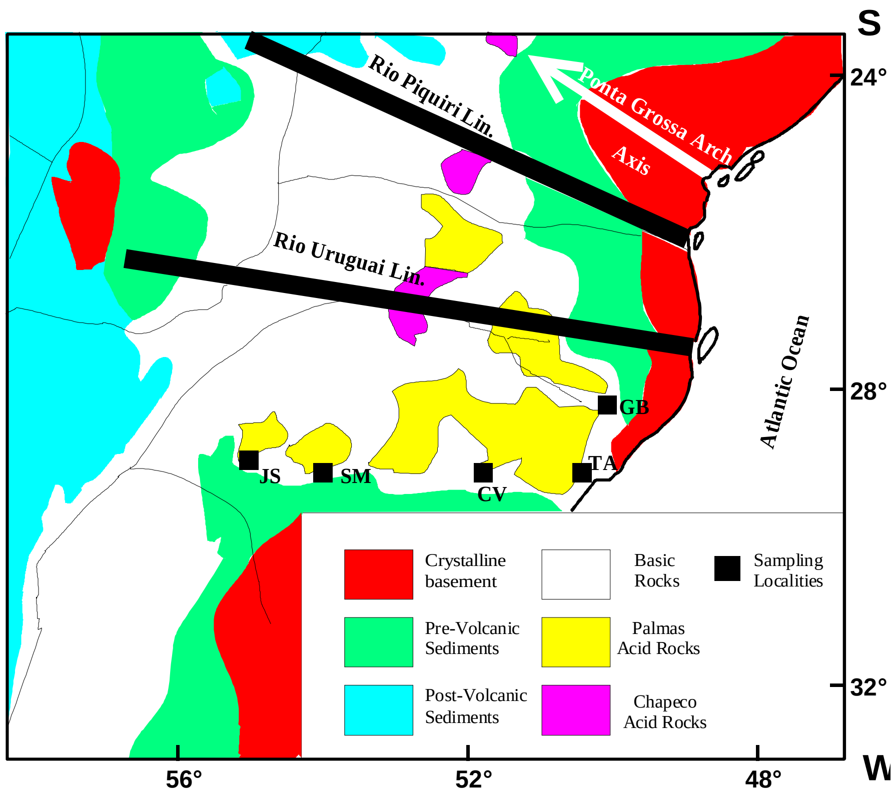

Figure 1. Simplified geological map of Paraná flood basalts with indication of paleomagnetic sections, modified from Kosterov et al. [1998].

Two sequences were sampled (35 lava flows, about 300 standard paleomagnetic cores). One sequence consist of 156 cores belonging to 18 lava flows collected at Salto Caixa hydroelectrical station and along the road Capitan Leonidas – Cascavel (SC: 25°24'S, 53°34'W). The second sequence 17 cooling sampled along the highway BR153 from Rio Uruguay level to Concordia and Irani (CON: between 27°22'S, 51°59'W and 26°47'S, 51°42'W). It is quite possible that our section SC corresponds to section CI from Tamrat et al. (1999) although no paleomagnetic holes were found at our sampling area.

Rock-Magnetic Properties of Selected Samples

Magnetic characteristics of typical samples selected for Thellier paleointensity measurements are summarized in Figure 2 and could be described as follows:

1. Selected samples carry essentially a single and stable component of magnetization, observed upon thermal (Figure 2a) treatment. A generally minor secondary component, probably of viscous origin, was present but was easily removed at either low temperature or low AF peak steps. The greater part of remanent magnetization, in most cases was removed at temperatures between 500 and 570°C, which points to low-Ti titanomagnetites as responsible for the magnetization.

2. Low-field continuous susceptibility measurements with temperature show the presence of a single ferrimagnetic phase with Curie point compatible with Ti-poor titanomagnetite (Figure 2b). Polished section observations under microscope also confirmed the presence of near magnetite phase associated with exsolved ilmenite of trellis or sometimes sandwich texture (Haggerty, 1976).

3. Hysteresis measurements at room temperature (using AGFM-Micromag apparatus) show (Figure 2c) that the studied samples fall in the ‘small’ pseudo-single-domain grain size region. This probably indicates a mixture of multidomain and a significant amount of single-domain (SD) grains.

In all we selected 117 samples for the paleointensity experiments, which belong to 19 lava flows having the above-described magnetic characteristics.

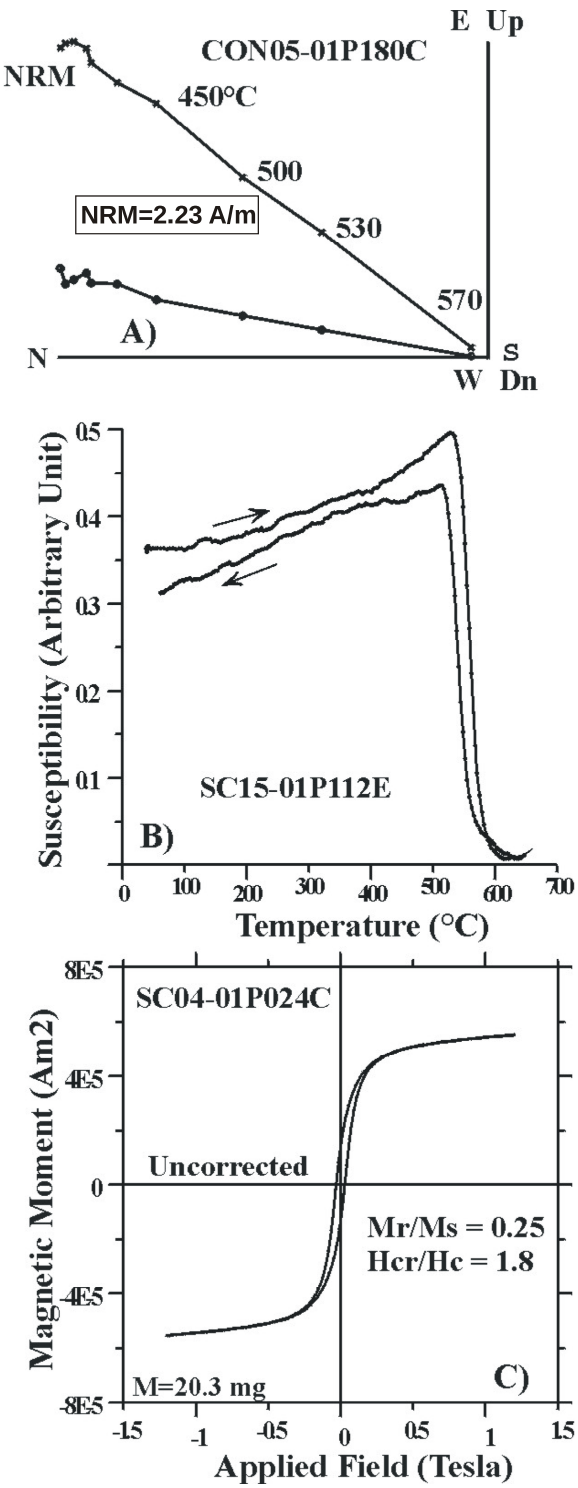

Figure 2. Summary of the magnetic characteristics of the typical samples, selected for Thellier paleointensity experiments. A) Orthogonal vector plot of stepwise thermal demagnetization (stratigraphic coordinates). The numbers refer to temperatures in °C. o – projections into the horizontal plane, x – projections into the vertical plane. B) Susceptibility versus temperature curve. The arrows indicate the heating and cooling curves. C) Example of hysteresis loop (uncorrected).

Paleointensity Experiments

Paleointensity experiments were performed using the Thellier method in its classic form (Thellier and Thellier, 1959). All heatings were made in a vacuum better than 10^{-2} mbar. The temperature settings were established from studies of the unblocking temperature spectrum of remanent magnetization (see above). Eleven temperature steps were distributed between room temperature and 560°C, and the laboratory field was set to 30 μT. Control heatings, commonly referred as pTRM checks, were performed after every heating step throughout the whole experiment (Figure 3). All remanences were measured using a JR5A spinner magnetometer.

Paleointensity data are reported on the classical Arai-Nagata (Nagata at al., 1963) plot on Figure 3, and results are given in Table 1. We accepted only determinations: (1) obtained from at least 5 NRM-TRM points corresponding to a NRM fraction larger than about 1/3 (Table 1), (2) yielding quality factor (Coe, et al., 1978) generally above 5, (3) with positive 'pTRM' checks i.e. the deviation of 'pTRM' checks were less than 15% and (4) with reasonably linear Zijderveld diagrams obtained from the paleointensity experiments.

Discussion and Main Results

Forty-two samples from 11 lava flows yielded apparently reliable absolute intensity determinations. The NRM fraction f used for paleointensity determination ranges between 0.30 to 0.74 and the quality factor q (Coe et al., 1978) varies from 4.3 to 17.3, being normally greater than 5 (Table 1). These results correspond to data of good technical quality. One lava flow (CON09) is represented by a single but high technical quality determination. This sample was omitted in calculating mean paleointensity or virtual dipole moment (VDM). Eleven comparable technical quality individual determinations (belonging to three southern Paraná cooling units) from Kosterov et al. (1998) data set are also incorporated in Table 1. However, site JS09, although yielded a high technical quality determination, is suspected to carry a chemical remanent magnetization (Kosterov, personal communication).

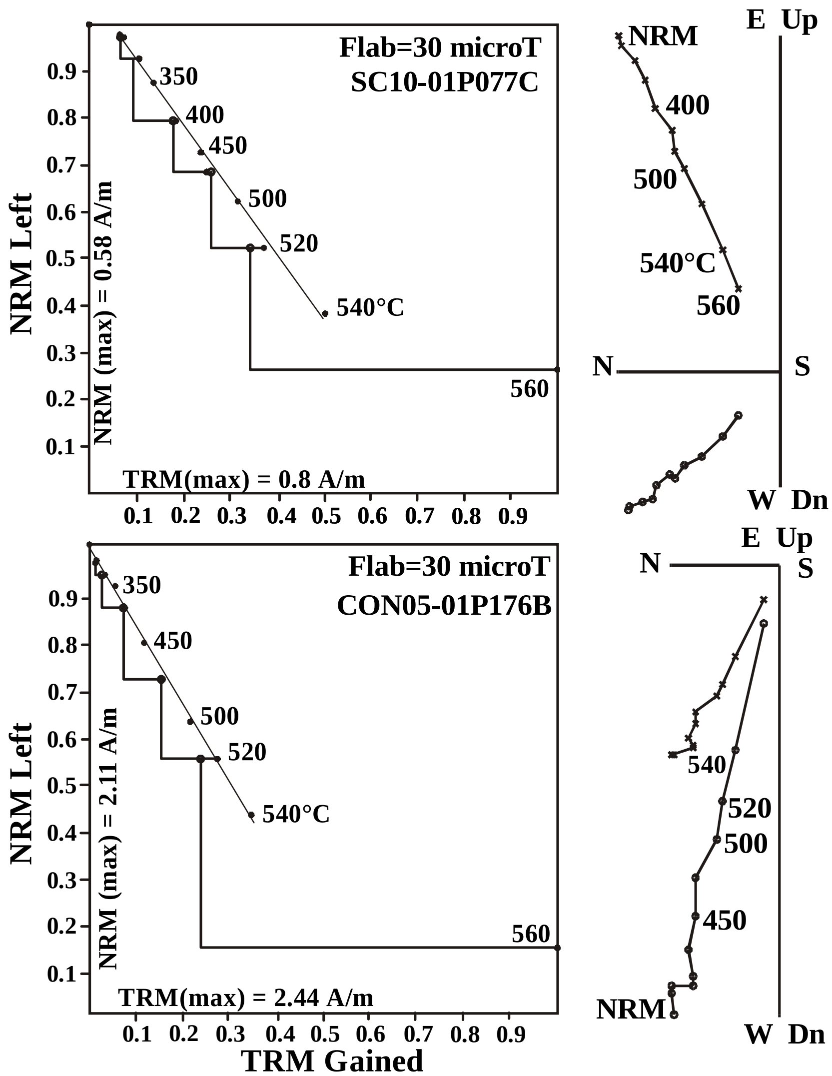

Figure 3. The representative NRM-TRM plots and associated orthogonal diagrams for Paraná samples. In the orthogonal diagrams we used same notations as in the figure 2a.

The mean paleointensity values per flow obtained in this study are ranging from 19.4 ± 4.0 to 46.7 ± 7.0 μT and the corresponding Virtual Dipole Moments are ranging from 4.0 ± 0.6 to 10.5 ± 1.1 (10^{22} Am^2). These data yield a mean value of 7.2 ± 2.3 x 10^{22} Am^2, which is about 92% of the present geomagnetic axial dipole (7.8 x 10^{22} Am^2 after Barton et al., 1996).

Paraná mean VDM is shown on Figure 4 together with 14 other selected mean VDMs and VADMs (virtual axial dipole moment) for the period 5-160 Ma. We selected data using quite modest criteria (Riisager et al., 2002) demanding a) the mean based on more than 9 successful determinations from at least three cooling units, b) no transitional data and c) paleointensity estimates obtained with Thellier method with pTRM checks. The perhaps most striking observation from inspecting Figure 4 is that, in spite of the many published paleointensity studies, there exists only 15 paleomagnetic dipole moment estimates in the 5-160 Ma period that fulfil the above mentioned modest criteria. Based on these data it is not possible to make firm conclusions about the evolution of geomagnetic intensity through geological time. Our results altogether with Tarduno et al's (2001) data suggest that the paleostrength during early Cretaceous period may be comparable or even much higher than recent field intensity and not 'anomalously low' as suggested by several authors. Thus, the proposal for the Mesozoic Dipole Low may probably need to be revised.

Table 1

Site	*Sample*	*Inc*	*Dec*	*n*	*Tmin-Tmax*	*f*	*g*	*q*	$F_E \pm \sigma(F_E)$	*VDM*	**F_E ± s.d.**	*VDMe*
SC01	01P002D	-46.7	3.6	7	300-520	0.66	0.79	8.8	21.5 ± 1.1	4.32	**24.6 ± 4.8**	**5.1 ± 0.9**
	01P003C	-40.6	8.7	6	350-520	0.64	0.74	5.6	20.8 ± 1.3	4.44		
	01P005B	-42.9	3.9	6	350-520	0.51	0.77	4.4	24.7 ± 2.6	5.16		
	01P007D	-47.2	358.9	8	250-520	0.59	0.81	10.4	31.4 ± 1.6	6.27		
SC04	01P024B	-45.8	6.1	8	300-540	0.58	0.82	6.3	19.7 ± 1.3	3.99	**19.4 ± 2.7**	**4.0 ± 0.6**
	01P029B	-40.5	15.3	8	300-540	0.56	0.85	9.8	21.9 ± 1.1	4.68		
	01P031D	-40.2	1.6	7	300-520	0.36	0.85	4.3	16.5 ± 1.0	3.39		
SC08	01P057A	-35.5	5.5	9	250-540	0.60	0.83	8.5	37.8 ± 2.6	8.45	**43.5 ± 5.9**	**9.5 ± 1.2**
	01P060B	-40.8	358.6	8	250-520	0.47	0.84	7.8	49.1 ± 2.5	10.5		
	01P061B	-38.5	355.8	8	250-520	0.50	0.82	6.7	48.1 ± 2.8	10.5		
	01P062D	-39.3	2.7	8	250-520	0.54	0.85	6.8	39.1 ± 2.5	8.45		

SC10	01P075D	-42.6	3.3	7	250-520	0.32	0.81	5.4	35.7 ± 2.3	7.48	**35.3 ± 5.7**	**7.4 ± 1.4**
	01P076B	-48.1	355.6	8	200-520	0.46	0.86	6.6	29.1 ± 1.7	5.75		
	01P077C	-45.1	1.9	9	200-540	0.65	0.85	12.2	31.2 ± 1.1	6.35		
	01P078C	-40.3	8.6	7	250-500	0.31	0.76	12.5	43.9 ± 0.9	9.4		
	01P079C	-41.1	2.5	7	250-500	0.39	0.82	4.4	36.4 ± 2.2	7.74		
SC15	01P110C	-25.3	13.8	10	20-540	0.46	0.88	6.8	22.9 ± 1.3	5.5	**22.1 ± 1.9**	**5.3 ± 0.5**
	01P111D	-28.1	11.3	8	250-520	0.42	0.84	6.5	24.3 ± 1.8	5.74		
	01P112C	-26.1	8.8	8	250-520	0.44	0.80	5.8	21.6 ± 1.4	5.16		
	01P113E	-23.3	5.9	7	300-520	0.51	0.81	4.9	21.3 ± 2.1	5.17		
	01P115B	-27.3	11.7	8	250-520	0.54	0.86	4.8	23.5 ± 2.3	5.58		
	01P116A	-29.7	11.6	7	250-500	0.30	0.79	6.1	19.1 ± 1.0	4.46		
CON03	01P159B	-54.3	10.1	10	20-540	0.65	0.87	12.5	38.4 ± 1.8	7.06	**33.4 ± 3.5**	**6.4 ± 0.5**
	01P161B	-54.4	7.7	9	250-560	0.74	0.86	17.3	32.8 ± 1.2	6.02		
	01P165C	-48.6	8.1	8	250-520	0.51	0.86	12.7	32.3 ± 1.5	6.35		
	01P167C	-47.5	5.4	9	250-540	0.55	0.84	7.3	30.2 ± 1.9	6.01		
CON05	01P176B	-35.1	8.1	10	20-540	0.58	0.86	12.4	45.5 ± 1.8	10.2	**45.6 ± 4.5**	**10.5 ± 1.1**
	01P178B	-31.1	10.2	9	200-520	0.39	0.85	4.6	40.1 ± 3.7	9.27		
	01P181B	-29.2	11.1	9	250-540	0.45	0.79	7.6	48.5 ± 3.8	11.4		
	01P182C	-30.5	9.2	9	250 - 540	0.64	0.83	5.9	42.1 ± 3.8	9.78		
	01P183C	-31.3	11.3	9	200-520	0.47	0.81	5.4	51.1 ± 3.2	11.8		
CON06	01P189C	-45.7	3.7	8	250-520	0.38	0.83	6.4	46.6 ± 2.6	9.46	**43.6 ± 2.7**	**8.8 ± 0.6**
	01P190B	-43.8	2.8	8	250-520	0.31	0.83	6.2	42.6 ± 1.7	8.82		
	01P191B	-48.5	5.3	9	20-520	0.47	0.89	13.3	41.5 ± 1.4	8.17		
CON07	01P192B	-48.1	6.2	8	250-520	0.33	0.84	3.8	43.2 ± 2.9	8.54	**46.7 ± 7.0**	**9.6 ± 1.4**
	01P194B	-45.6	6.8	8	250-520	0.34	0.83	5.3	54.9 ± 3.1	11.2		
	01P195C	-44.6	9.6	8	250-520	0.35	0.81	7.8	49.9 ± 1.8	10.2		
	01P199B	-40.7	7.6	9	250-540	0.39	0.82	9.8	39.1 ± 1.8	8.34		
CON09	01P210B	-36.8	8.2	8	250-520	0.51	0.81	7.9	26.2 ± 1.4	5.79		
CON14	01P247B	-43.2	350.5	8	250-520	0.42	0.84	7.6	31.1 ± 1.4	6.48	**27.3 ± 6.4**	**5.6 ± 1.3**
	01P251B	-45.6	347.3	8	250-540	0.64	0.0.84	4.5	30.8 ± 3.7	6.26		
	01P253B	-45.1	352.8	9	200-540	0.68	0.88	11.8	19.9 ± 1.1	4.06		

Table 1. Paleointensity results from Parana volcanic units, Inc and Dec are magnetic inclination and declination of cleaned remanence of individual samples, n is number of NRM-TRM points used for palaeointensity determination, Tmin-Tmax is the temperature interval used, *f*, *g* and *q* are the fraction of extrapolated NRM used, the gap factor and quality factor (Coe et al., 1978) respectively. F_E is paleointensity estimate for individual specimen, and $\sigma(F_E)$ is its standard error; F_E is average paleointensity of individual lava flow, the plus and minus sign corresponding to standard deviation; VDM and VDMe are individual and average virtual dipole moments. (*) Site JS09, although yielded a high technical quality determination, is suspected to carry a chemical remanent magnetization (Kosterov, personal communication).

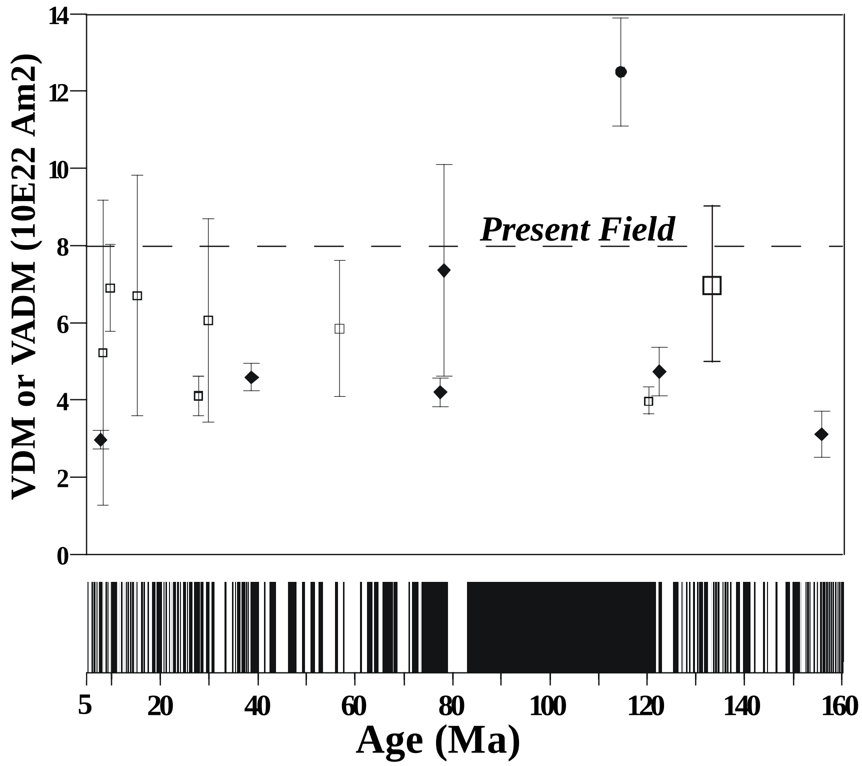

Figure 4. Evolution of mean virtual dipole moments (VDMs) and virtual axial dipole moments (VADMs) for 5 to 160 Ma. Open squares are mean VDM obtained from continental lava flows (Goguitchaichvili et al., 2000, 2001; Prévot et al., 1985; Riisager et al., 1999, 2000, 2002; Zhu et al., 2001) (This study – big symbols). Diamonds are mean VADMs obtained from submarine basaltic glass (Selkin and Tauxe, 2000). Closed circle is mean VDM obtained single plagioclase crystals (Tarduno et al., 2001). Also shown is the geomagnetic polarity time scale to 160 Ma (Cande and Kent, 1995).

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Conclusiones Generales

Una identificación precisa de los portadores de la magnetización, así como de su estado de dominio magnético, es de fundamental importancia para la determinación de paleointensidades, ya que 1) el significado geomagnético de una determinación depende de que la magnetización natural sea original y proporcional a la intensidad del campo magnético al momento de su formación y 2) solo la remanencia portada por granos de dominio sencillo (SD) obedecen las leyes de las magnetizaciones termorremanentes parciales de Thellier, premisa fundamental para la aplicación del método de paleointensidades que lleva su nombre [Thellier & Thellier,1959].

Las curvas termomagnéticas continuas permiten la determinación de puntos de Curie de minerales magnéticos y estimar su estabilidad térmica. Así, ellas son herramientas indispensables en la identificación de los portadores de la magnetización y en la selección de las muestras más adecuadas para experimentos de paleointensidad absoluta. Tres tipos de curvas termomagnéticas son usados rutinariamente en paleomagnetismo: (1) Magnetización inducida (saturación) versus temperatura, conocidas como curvas J_s-T, (2) Susceptibilidad versus temperatura, o curvas k-T y (3) Magnetización remanente vs temperatura, o curvas J_{rs}-T. No existe aún acuerdo entre la comunidad paleomagnética sobre cuál método es más apropiado y sensitivo, especialmente en casos donde se encuentran presentes dos o más fases magnéticas. A fin de clarificar esta cuestión se realizó una investigación termomagnética comparativa obteniendo curvas continuas k-T y J_s-T sobre las mismas muestras volcánicas naturales. En algunos casos se obtuvieron también curvas J_{rs}-T utilizando un termo-magnetómetro de vibración (VSTM). En cualesquiera de los caso analizados, las curvas k-T parecen ser más sensitivas para la evolución térmicas de muestras naturales. Lo anterior, además, es apoyado por inferencias teóricas. Así, una determinación precisa de la mineralogía magnética se puede obtener de las curvas k-T.

La determinación del estado de dominio magnético es importante en paleomagnetismo porque la estabilidad de la señal magnética depende de la estructura de dominio de los minerales opacos. La identificación anticipada de granos de dominio múltiple o granos de pseudo dominio sencillo “grandes” es decisiva en los estudios de paleointensidad absoluta, ya que ellos violan las leyes de Thellier de las magnetizaciones termoremanentes y no se pueden obtener paleointensidades geomagnéticas de este tipo de materiales. Los diagramas de Day son utilizados rutinariamente en artículos de investigación en paleomagnetismo y en magnetismo de rocas. Sin embargo, las rocas casi siempre muestran un comportamiento del tipo pseudo dominio sencillo, según sus parámetros de histéresis. La estructura de dominio magnético puede estimarse de forma diferente. Ya que las temperaturas de bloqueo y desbloqueo de granos de dominio múltiple son diferentes, la presencia de dichos granos puede ser detectada por medio de experimentos de adquisición de magnetizaciones termoremanentes parciales (pTRM) y su desmagnetización posterior. Al aplicar ambos métodos a las mismas muestra volcánicas vírgenes, las cuales contienen “casi magnetita pura”, como lo mostraron estudios previos, los métodos termomagnéticos fueron más resolutivos que las mediciones de histéresis regulares.

La confiabilidad de muchas determinaciones de paleointensidad puede ser cuestionable, dependiendo de la metodología empleada. Pero ¿como se pueden reconocer determinaciones de PI confiables?

Las rocas volcánicas mas viejas que unos cuantos millones de años son más susceptibles a la alteración durante los calentamientos de laboratorio y se debe tener cuidado de no interpretar tal alteración en términos geomagnéticos. Lo anterior esta bien ilustrado en un trabajo de Goguitchaichvili et al, 1999 sobre transiciones paleomagnéticas en Islandia, en donde se mostró que la gran dispersión presentada en los datos de PI obtenidos en estudios previos, realizados sin la inclusión de ‘pTRM’s checks’, estaba causada por alteración no detectada durante los calentamientos de laboratorio. Otras varias causas relacionadas a experimentos de PI en rocas volcánicas son discutidas en (Riisager et al., 2002 y Calvo et al., 2002.

Es importante hacer notar que en la compilación más reciente de paleointensidades globales, aproximadamente 30 % de éstas son determinaciones Rusas (Perrin y Shcherbakov, 1997) obtenidas por el método de Thellier y Thellier sin 'pTRM's checks', publicados desde los años 60's. Es muy difícil, bajo estas circunstancias, estimar la confiabilidad de estos datos en ausencia de cierta información geológica y magnética crucial. Una característica particular de todos los datos de PI publicados por los colegas rusos es la notable baja variación paleosecular entre unidades de enfriamiento independientes; lo cual sugiere ya sea, una variación paleosecular inusualmente baja en la ex Unión Soviética, comparada contra otras localidades geográficas, o bien, la existencia de diferencias experimentales significativas entre los estudios rusos y los desarrollados en cualquier otro lado.

Para mostrar el hecho anterior, baste mencionar que las PI obtenidas de 33 unidades de enfriamiento Eocénicas (Solodovnikov, 1998) estan extremadamente bien agrupadas con una desviación estándar de 16 % (Figura 1a). Una selección de datos globales de PI entre 0 y 5 Ma presenta una clara distribución log-normal (Riisager, 1999) con una desviación estándar de alrededor de 42% (Figura 1c). Algunos datos rusos más antiguos poseen desviaciones estándares aún más bajas!. Así, es muy posible que estos datos no representen características reales del campo geomagnético.

De igual forma, aquellas determinaciones obtenidas mediante la técnica propuesta por Shaw (1974) y sus modificaciones (Kono, 1978; Rolph y Shaw, 1985; podrían ser cuestionables. Un estudio de PI mediante el método de Shaw original, efectuado en rocas volcánicas del Paleoceno, reportó una desviación estándar inusualmente alta de 67% (Figura 1b) sin poseer una distribución log-normal.

Una evaluación experimental del método de Shaw original y sus variantes (capitulo II del presente trabajo; ver también Morales et al., 2003) mostró que estos experimentos pueden sobreestimar la intensidad del campo magnético terrestre. Más del 75 % de los datos disponibles para el intervalo que va de los 5 a los 160 Ma fueron obtenidos mediante el método de Shaw o Thellier sin pTRM's check's y por tanto, deben ser tratados con extremo cuidado por las razones antes mencionadas.

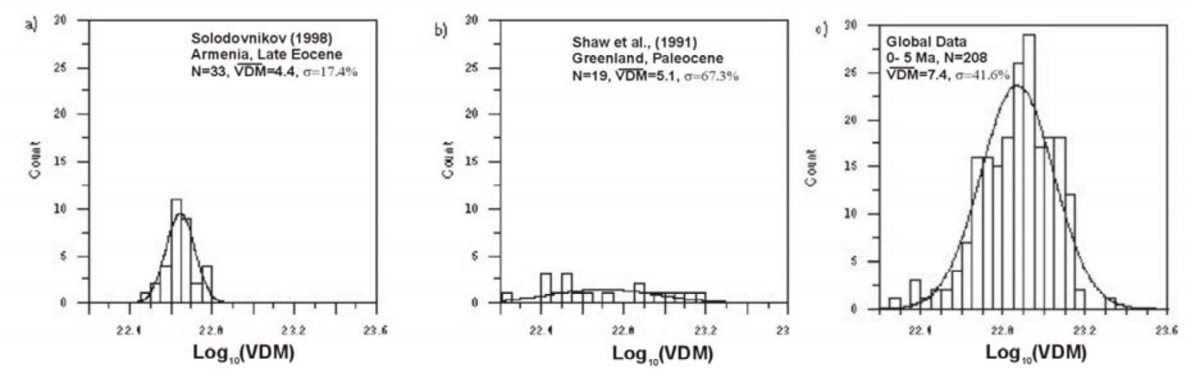

Figura 1. (Distribución log-normal de los momentos dipolares virtuales basados a) Slodonikov et al., b) Shaw et al 1991 y c) datos globales entre 0 y 5 Ma).

El empleo de la técnica de microondas en los experimentos de PI parece muy atractivo y eficaz, siempre y cuando se tomen en cuenta aquellos factores que podrían dar lugar a resultados poco confiables. Sin embargo, esta metodología merece aún mas trabajo de investigación antes de que se puede concluir que "la tasa de éxito" de experimentos tipo Thellier empleando microondas sea equivalente, o inclusive mayores que la de experimentos de Thellier convencionales (Biggin y Thomas, 2003)". La mejor forma, tal vez, de confirmar la confiabilidad de una metodología dada es mediante la utilización de flujos de lava históricos, cuya intensidad del campo magnético antiguo puede ser bien conocida. Con esta idea en mente se realizó un estudio de paleointensidad de microondas en flujos del volcán Paricutín.

Las muestras seleccionadas dieron resultados significativamente diferentes de la intensidad esperada. La dispersión observada en los datos obtenidos por ésta técnica sugieren la ocurrencia de efectos internos en los flujos de lava como posible causa de tal dispersión, o bien, puede deberse al hecho de que solo una pequeña fracción de la magnetización remanente fue utilizada para la estimación de la paleointensidad. Son necesarios experimentos adicionales empleando microondas de mayor frecuencia (14 GHz) a fin de discernir las causas del fracaso de los experimentos realizados.

Se presentaron los resultados de una investigación realizada en rocas volcánicas mexicanas miocénicas, para las cuales se obtuvieron las primeras determinaciones confiables de paleointensidad para Baja California, habiendo obtenido resultados de excelente calidad técnica. El valor medio del momento dipolar virtual para las rocas miocénicas de Baja California, obtenido en este estudio, resultó ser aproximadamente el 85% del valor del dipolo axial geomagnético presente, en concordancia, también, con aquel valor reportado para rocas volcánicas Miocénico-tardías de México Central (Goguitchaichvili et al., 2000a) y con aquel para las basaltos de Steens Mountain. A pesar de la abundancia relativamente grande de datos para el periodo Plio-Pleistoceno, su dispersión grande y su baja distribución de datos confiables hacen difícil la formulación de conclusiones firmes sobre la evolución temporal del campo magnético.

A fin de contribuir a disminuir la disparidad existente en el número de determinaciones de PI entre los dos hemisferio [Pick y Tauxe, 1993; Juarez y Tauxe, 2000], la cual impide un análisis fino de los cambios en los parámetros que caracterizan al campo magnético, se prestó particular interés en obtener determinaciones de PI de formaciones localizadas en el hemisferio sur. Para ello se realizaron estudios de PI en formaciones de la parte norte de Chile (Formación la Negra) y de flujos de Paraná (Brasil). La buena calidad técnica de los resultados obtenidos, aunque no numerosos, aunada a la aplicación de un criterio de selección de resultados estricto, nos permitió darnos cuenta de que a pesar del gran número de estudios de PI publicados solo 17 estimaciones del momento dipolar geomagnético para el periodo 5 a 200 Ma cumplen con dicho criterio estricto de selección de resultados.

Estas dos formaciones son de particular importancia no solo por su distribución geográfica sino también porque permitieron cuestionar la validez del así conocido como “bajo dipolar del Mesozoico”; implicación propuesta primeramente por Prévot et al., 1990 y que asigna una intensidad de campo magnético para este periodo de tan solo una tercera parte del valor actual.

Las dificultades relacionadas con los experimentos de paleointensidades de Thellier (1959) son evidentes al revisar los estudios mas actuales citados. Porcentajes altos de experimentos fallidos no son sorprendentes ni exclusivos del método de Thellier, cuando se realiza una adecuada preselección de las muestras a estudiar y cuando se realiza un análisis estricto de los datos obtenidos. El método de Thellier (1959), el cual es considerado el más confiable (Goguitchaichvili et al., 1999; Juarez y Tauxe 2000; Cottrel y Tarduno, 2001 y Scherbakov, et al., 2001), impone muchas restricciones para la selección de muestras que pueden ser usadas para determinaciones exitosas (Coe, 1967; Levi, 1977; Prévot et al., 1985; Pick y Tauxe, 1993a, Kosterov y Prévot, 1998).

Factores de magnetismo de rocas son mas serios de lo que parecen en estudios de paleointensidades como para ser reconocidos, especialmente aquellos asociados con cambios químicos que se presentan en la naturaleza pero no en el laboratorio, i. e., magnetizaciones químicas remanentes en vidrios basálticos submarinos (SBG).

Tres tipos diferentes de variaciones de la paleointensidad en tiempo geológico han sido visualizados por Heller et al., 2002 (Figura 2). El modelo 1 sugiere una paleointensidad baja para todo el mesozoico. El modelo 2, que contiene principalmente datos obtenidos de vidrios basálticos submarinos (Selkin y Tauxe, 2000; Juarez et al., 2000), apoya la idea de paleointensidades relativamente altas a partir de 0.3 Ma con valores previos mas bajos a aquellos que incluye al super chron normal cretáceo (CNS). Por el contrario, Tarduno et al. (2001, 2002) sugiere que la paleointensidad fue mas alta durante el CNC (tendencia 3). Sin embargo, tomando en cuenta que solo 17 momentos dioplares promediados en el tiempo pueden ser considerados como confiables

(Goguitchaichvili et al, 2003), es singularmente difícil formular alguna conclusión firme sobre las variaciones a gran escala de la intensidad del campo magnético. Por el contrario, se observa que estos datos de alta calidad no parecen ajustarse a ninguno de los modelos propuestos y que ni siquiera muestran alguna variación sistemática.

La existencia del bajo dipolar para el mesozoico es aún cuestionable. Históricamente, bajo este término se ha considerado un intervalo bastante grande (180 – 135 Ma). McElhinny y Larson, 2003 mostraron recientemente el mínimo de la intensidad del campo geomagnético para el intervalo de 150 – 167 Ma denominado Jurasic Quite Zone (JQZ por sus siglas en ingles). Los datos oceánicos obtenidos a partir de las amplitudes de las anomalías magnéticas concuerdan perfectamente con aquellos datos continentales (Biggin y Thomas, 2002) para el mismo intervalo. Estos autores proponen restringir el bajo dipolar para el mesozoico únicamente al JQZ. Cabe mencionar que nuestros datos seleccionados apoyan la hipótesis sugerida ya que el único dato de que disponemos para dicho intervalo concuerda bastante bien con la baja intensidad propuesta, con un valor del momento dipolar axial inferior a $3*10^{22}$ Am^2 (ver Figura 2).

Son indispensables más datos confiables antes de poder establecer cualquier relación entre la paleointensidad y los procesos geodinámicos globales.

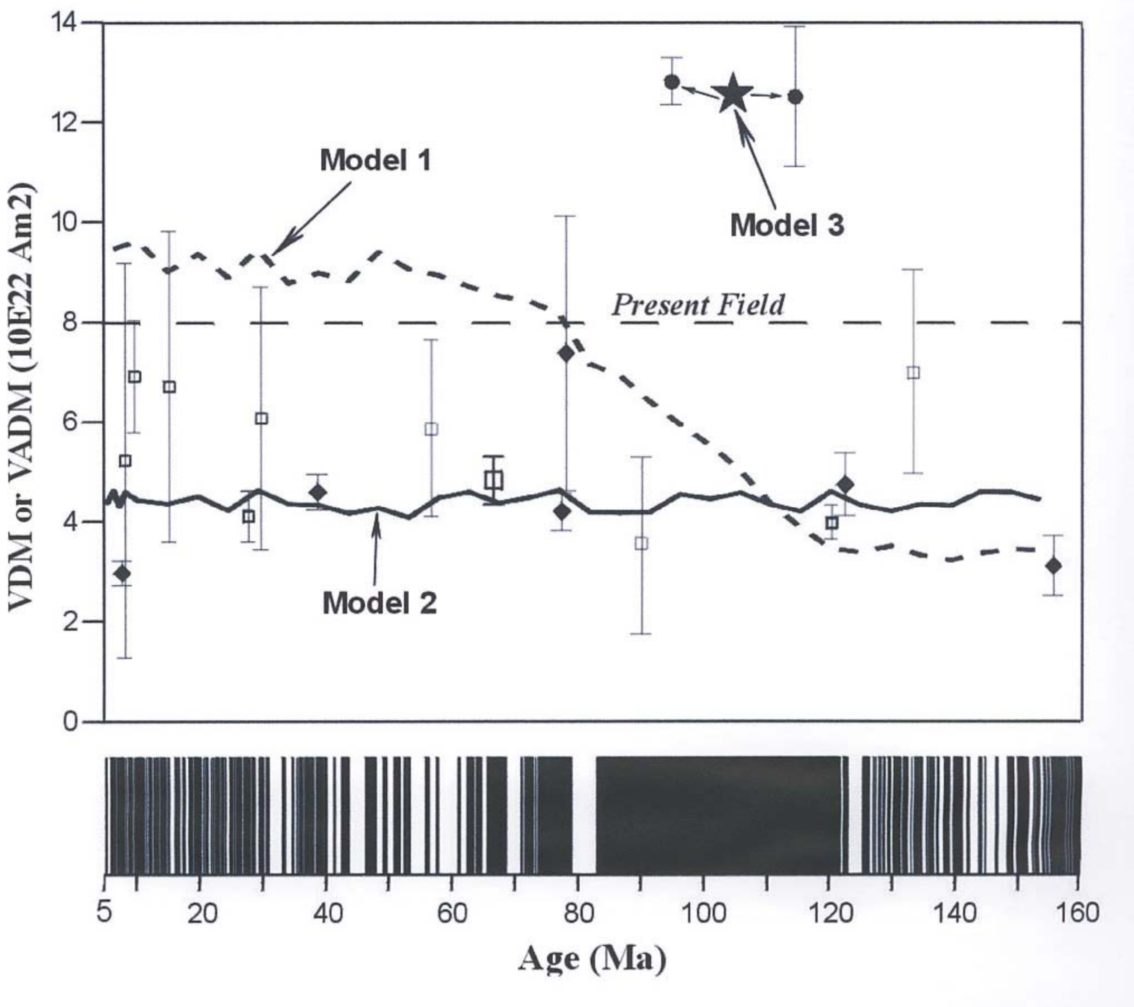

Figura 2. Diagrama en el que se muestran diferentes modelos para la peleointensidad media. El modelo 1 (línea discontinua) tiene una intensidad media durante parte del mesozoico la cual es significativamente mas baja que aquella durante el Cenozoico (Prévot et al., 1990), el modelo 2 (línea sólida) muestra una intensidad media durante los últimos 300 kyr que es casi dos veces mayor que la correspondiente a los 160 Ma precedentes (Selkin yT auxe, 2000), mientras que el modelo 3 (estrella) sugiere una intensidad media alta durante el CNS (Tarduno et al., 2001). Tomado de Heller et al., 2002.

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