Palaeointensities determined from the middle Cretaceous basalt in Liaoning Province, northeastern China

Rixiang Zhu^{a,*}, Ching-Hua Lo^b, Ruiping Shi^a, Guanghai Shi^a,
Yongxin Pan^a, Jian Shao^c

^a *Institute of Geology and Geophysics, Chinese Academy of Sciences, P.O. Box 9825, Beijing 100101, China*

^b *Department of Geosciences, National Taiwan University, P.O. Box 9825, 245 Choushan Road, Taipei 106, China*

^c *Department of Geology, Peking University, P.O. Box 9825, Beijing 100871, China*

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Abstract

A palaeomagnetic and geochronologic investigation has been conducted on a basalt lava sequence at Jianguo in Liaoning Province, northeastern China. Radiometric $^{40}\text{Ar}/^{39}\text{Ar}$ dating experiments indicate the volcanism occurred in relatively short time intervals between 105.5 and 102.2 Ma. Thermomagnetic experiments and electron microprobe show that main primary carrier of remanence to be titanomagnetite with Curie temperature of $\sim 300^\circ\text{C}$. A modified version of the Thellier–Thellier technique with systematic partial thermoremanent magnetization (pTRM)-checks was used for palaeointensity determinations. Fourteen out of 61 samples from the studied lavas yield reliable palaeointensity estimates with a mean virtual dipole moment (VDM*) of $(4.07 \pm 0.2) \times 10^{22} \text{ Am}^2$, about half the strength of the modern-day field.

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1. Introduction

The behavior of the geomagnetic field during the Cretaceous normal superchron (CNS) is critical for understanding the geodynamo within the Earth. Abnormally high palaeointensity within the CNS has been suggested based on the evidence that the mantle plume activity was higher in the Cretaceous (Larson and Olson, 1991). This idea was supported by computation simulation and palaeointensity determinations using single plagioclase crystals (Glatzmaier et al.,

1999; Tarduno et al., 2001, 2002). Conversely, low palaeointensity within the CNS has been reported with the Thellier–Thellier technique using submarine basaltic glass or bulk continent basalt (Pick and Tauxe, 1993; Juarez et al., 1998; Selkin and Tauxe, 2000; Sherwood et al., 1993; Tanaka and Kono, 2002). Therefore, more reliable palaeointensity estimates with precise ages are needed before we can conclude whether the average field strength during the CNS was significantly stronger or weaker than that out of it. Given this global incongruence of palaeointensity data during the CNS, we present new palaeointensities determined from Cretaceous continental basalt in Liaoning Province, northeastern China.

* Corresponding author. Tel.: +86-10-62007912;
fax: +86-10-62010846.
E-mail address: rxzhu@mail.igcas.ac.cn (R. Zhu).

2. Geological setting and dating

2.1. Geological setting

The Jianguo section ($41^{\circ}16'N$, $121^{\circ}54.2'E$) is situated the northern part of the Fuxin–Yixian basin in Wester Liaoning (Fig. 1). This basin has attracted the attention of geologists and paleontologists because of its well-preserved and unique early Cretaceous fossil assemblages, known as Jehol fauna, in the Yixian Formation strata (Pan et al., 2001; Zhou et al., 2003). Geologically, the Jianguo section is composed of the upper part of the Dalinghe (DLH) Formation, which mainly consists of grey-purplish andesites, pyroxene-andesite lava and blackish basalts (Bureau of Geology and Mineral Resources of Liaoning, 1970). A recent study on the lower andesites indicated that the latitude of the studied area has no obvious changes since the DLH Formation formed (Zhu et al., 2004) though it is not necessarily free of regional deformation.

No sedimentary horizons have been observed between basaltic lava flows. Through vesicular zones and scoria, five lava flows (lava #1–5) were recognized in field with thickness ranging from 5 to 10 m. Using a portable gasoline-powered drill, a total of 153 orientated cores were collected, no less than twenty fresh cores per flow. The samples were orientated in situ using a sun compass and/or a magnetic compass, while two fresh massive samples (sample JG-1 from the uppermost unit and JG-5 from the lowest unit of the volcanic sequence) were collected for radiometric dating.

2.2. Radiometric dating

Two whole-rock samples JG-1 and JG-5 were dated by $^{40}\text{Ar}/^{39}\text{Ar}$ laser step-heating method. Samples were first crushed and disintegrated. After sieving, mineral grains and rock chips in the range of 140–250 μm were ultrasonically cleaned in distilled water and dried, and then handpicked to remove

Fig. 1. Schematic map showing Cretaceous rock distribution in studied area and the sampling site of Jianguo remarked with star symbol (modified from Zhu et al., 2003b).

visible contamination. The samples were then irradiated together with the LP-6 biotite standard (Odin et al., 1982; Baksi et al., 1996) in the VT-C position at the THOR Reactor in Taiwan for 30 h. After irradiation, standards and samples were either incrementally heated using a US LASER Nd-YAG laser operated in continuous mode, and the gas was measured by a VG3600 mass spectrometer at the National Taiwan University. Ages were calculated from Ar isotopic ratios measured after corrections made for mass discrimination, interfering nuclear reactions, procedural blanks, and atmospheric Ar contamination. Detail analysis and data processing procedures were outlined

by Lo et al. (2002). The results of the $^{40}\text{Ar}/^{39}\text{Ar}$ experiments are shown in Table 1 and plotted as age spectra in Fig. 2.

As shown in Fig. 2, both samples exhibit a flat profile for the age spectra, with plateau age of 102.2 ± 0.5 and 105.5 ± 0.5 Ma for JG-1 and JG-5, respectively. These plateau ages are all consistent with their respective intercept ages obtained from the isotope correlation diagrams, 102.6 ± 1.3 for JG-1 and 105.4 ± 0.7 Ma for JG-5, respectively. Regressions of the data on isotope correlation diagrams indicate that the non-radiogenic argon component in the samples are mainly atmospheric, as indicated by

Table 1
Results of $^{40}\text{Ar}/^{39}\text{Ar}$ laser step-heating experiments

E^a	Atmospheric (%)	$^{36}\text{Ar}/^{39}\text{Ar}$	$^{37}\text{Ar}/^{39}\text{Ar}$	$^{38}\text{Ar}/^{39}\text{Ar}$	$^{40}\text{Ar}/^{39}\text{Ar}$	$^{40}\text{Ar}/^{36}\text{Ar}$	Age (Ma)
JG-1 whole rock							
0.23	10.07	0.74757E-02	0.22006E+01	0.27484E-01	0.20313E+02	0.27172E+04	100.6 $\pm$ 0.6
0.30	23.25	0.19285E-01	0.61655E+00	0.19786E-01	0.24335E+02	0.12618E+04	102.7 $\pm$ 1.3
0.42	16.47	0.12671E-01	0.10401E+01	0.24162E-01	0.22292E+02	0.17592E+04	102.4 $\pm$ 0.7
0.48	15.18	0.11499E-01	0.12178E+01	0.25027E-01	0.21814E+02	0.18971E+04	101.8 $\pm$ 0.7
0.57	13.16	0.97743E-02	0.14559E+01	0.25289E-01	0.21150E+02	0.21639E+04	101.1 $\pm$ 0.6
0.67	12.84	0.98816E-02	0.20058E+01	0.25639E-01	0.21601E+02	0.21860E+04	103.6 $\pm$ 0.8
0.74	13.52	0.10228E-01	0.19271E+01	0.27023E-01	0.21310E+02	0.20835E+04	101.5 $\pm$ 0.7
0.84	12.62	0.95254E-02	0.20342E+01	0.26120E-01	0.21119E+02	0.22171E+04	101.6 $\pm$ 0.6
0.93	11.37	0.84702E-02	0.18707E+01	0.26159E-01	0.20809E+02	0.24567E+04	101.5 $\pm$ 0.8
1.04	9.83	0.75356E-02	0.27934E+01	0.29604E-01	0.20530E+02	0.27244E+04	101.9 $\pm$ 0.7
1.15	10.42	0.78644E-02	0.22108E+01	0.24461E-01	0.20739E+02	0.26371E+04	102.3 $\pm$ 1.2
1.24	6.28	0.49558E-02	0.27373E+01	0.26179E-01	0.20078E+02	0.40515E+04	103.6 $\pm$ 0.9
1.31	3.39	0.30231E-02	0.29771E+01	0.27476E-01	0.19760E+02	0.65364E+04	105.0 $\pm$ 0.8
1.40	6.71	0.54199E-02	0.29613E+01	0.26079E-01	0.20559E+02	0.37933E+04	105.5 $\pm$ 0.9
1.50	6.13	0.51216E-02	0.35670E+01	0.27762E-01	0.20345E+02	0.39725E+04	105.1 $\pm$ 1.0
1.58	7.54	0.59737E-02	0.30273E+01	0.27128E-01	0.20421E+02	0.34185E+04	103.9 $\pm$ 1.4
1.68	8.21	0.67070E-02	0.40048E+01	0.26639E-01	0.20499E+02	0.30564E+04	103.7 $\pm$ 3.8
1.76	17.71	0.14239E-01	0.80728E+00	0.20479E-01	0.23442E+02	0.16463E+04	106.0 $\pm$ 1.3
1.84	10.96	0.85733E-02	0.30958E+01	0.25720E-01	0.21021E+02	0.24519E+04	103.1 $\pm$ 8.5
J value = $0.00313916 \pm 0.00001318$							
Total gas age = 102.2 ± 0.5 Ma							
Plateau age = 102.2 ± 0.5 Ma (0.23–1.84 W)							
JG-5 whole rock							
0.07	13.09	0.10071E-01	0.10142E+01	0.70423E-01	0.22182E+02	0.22025E+04	105.9 $\pm$ 1.2
0.24	24.48	0.21297E-01	0.12068E+01	0.84864E-01	0.25371E+02	0.11913E+04	105.3 $\pm$ 3.3
0.33	12.85	0.97928E-02	0.81153E+00	0.57711E-01	0.22066E+02	0.22533E+04	105.7 $\pm$ 0.9
0.43	8.75	0.64368E-02	0.10763E+01	0.47632E-01	0.20838E+02	0.32374E+04	104.5 $\pm$ 0.5
0.48	6.87	0.49979E-02	0.10195E+01	0.41949E-01	0.20410E+02	0.40836E+04	104.5 $\pm$ 1.2
0.54	5.39	0.41787E-02	0.19148E+01	0.43802E-01	0.20257E+02	0.48476E+04	105.4 $\pm$ 1.0
0.58	5.09	0.40710E-02	0.22640E+01	0.42861E-01	0.20327E+02	0.49932E+04	106.1 $\pm$ 0.9
0.64	6.83	0.53773E-02	0.24040E+01	0.40608E-01	0.20638E+02	0.38380E+04	105.8 $\pm$ 2.5
0.68	2.60	0.24945E-02	0.29075E+01	0.39691E-01	0.19922E+02	0.79863E+04	106.7 $\pm$ 1.5
0.72	6.00	0.48231E-02	0.26102E+01	0.40289E-01	0.20503E+02	0.42509E+04	106.0 $\pm$ 1.3
0.78	6.21	0.51688E-02	0.33587E+01	0.39041E-01	0.20556E+02	0.39769E+04	106.1 $\pm$ 1.8

Table 1 (Continued)

E^a	Atmospheric (%)	$^{36}\text{Ar}/^{39}\text{Ar}$	$^{37}\text{Ar}/^{39}\text{Ar}$	$^{38}\text{Ar}/^{39}\text{Ar}$	$^{40}\text{Ar}/^{39}\text{Ar}$	$^{40}\text{Ar}/^{36}\text{Ar}$	Age (Ma)
0.84	2.83	0.28058E–02	0.35931E+01	0.36611E–01	0.19756E+02	0.70411E+04	105.7±1.6
0.90	2.84	0.27732E–02	0.32377E+01	0.39194E–01	0.20289E+02	0.73162E+04	108.4± 2.2
0.98	3.78	0.34531E–02	0.36067E+01	0.42458E–01	0.19827E+02	0.57419E+04	105.0±1.1
1.06	4.78	0.40188E–02	0.30010E+01	0.38743E–01	0.20144E+02	0.50125E+04	105.5±1.4
1.12	3.93	0.35656E–02	0.34939E+01	0.37137E–01	0.20130E+02	0.56458E+04	106.4±1.8
1.20	3.80	0.35126E–02	0.35647E+01	0.40940E–01	0.20271E+02	0.57710E+04	107.3±4.2
J value = 0.00313916 ± 0.00001318							
Total gas age = 105.5 ± 0.5 Ma							
Plateau age = 105.5 ± 0.5 Ma (0.07–1.20 W)							

Note: J value is the weighted mean of three fusions of irradiation standard LP-6 Biotite having a calibrated $^{40}\text{Ar}/^{39}\text{Ar}$ age of 128.4±0.2 Ma, based on Fish Canyon Sanidine (28.02 ± 0.28 Ma) (Baksi et al., 1996; Renne et al., 1998).

The age is obtained by using the following equations:

$$\text{date} = \frac{1}{\lambda} \ln \left(1 + J \frac{^{40}\text{Ar}^*}{^{39}\text{Ar}_K} \right),$$

and

$$\frac{^{40}\text{Ar}^*}{^{39}\text{Ar}_K} = \frac{[^{40}\text{Ar}/^{39}\text{Ar}]_m - 295.5[^{36}\text{Ar}/^{39}\text{Ar}]_m + 295.5[^{36}\text{Ar}/^{37}\text{Ar}]_{\text{Ca}}[^{37}\text{Ar}/^{39}\text{Ar}]_m}{1 - ([^{39}\text{Ar}/^{37}\text{Ar}]_{\text{Ca}}[^{37}\text{Ar}/^{39}\text{Ar}]_m)} - \left[\frac{^{40}\text{Ar}}{^{39}\text{Ar}} \right]_K$$

where $[]_{\text{Ca}}$ and $[]_K$ are the isotope ratios of argon extracted from irradiated calcium and potassium salts (values cited in the text) and $[]_m$ is the isotope ratio of argon extracted from irradiated unknown.

Age (Ma) is the age calculated using the following decay constants: $\lambda_{\text{e}} = 0.581 \times 10^{-10}$ per year; $\lambda_{\text{B}} = 4.962 \times 10^{-10}$ per year; $\lambda = 5.543 \times 10^{-10}$ per year; $^{40}\text{K}/\text{K} = 0.01167$ at.% (Steiger and Jäger, 1977).

The quoted error is one standard deviation.

^a Laser energy in watt.

the obtained $^{40}\text{Ar}/^{39}\text{Ar}$ intercept values (286.9±21.4 with MSWD = 2.151 for JG-1 and 288.1 ± 14.9 with MSWD = 0.870 for JG-5) which are generally consistent with the atmospheric composition. These would suggest that the obtained plateau ages should be considered as the estimates of the eruption ages for the lava flows, which may have occurred in the period of 105.5–102.2 Ma.

3. Rock-magnetic investigations and Thellier's experiments

3.1. Rock-magnetic investigations

Magnetic properties were investigated on samples from each lava flow in order to identify the magnetic carriers of remanence and guide sample selection for palaeointensity experiments. All rock-magnetic investigations were conducted at the Palaeomagnetism Laboratory of the Institute of Geology and Geophysics, Chinese Academy of Sciences (IGG-CAS).

Thermomagnetic curves (J – T) of small amounts of powdered samples from each flow were measured in argon atmosphere on a Magnetic Measurements Variable Field Translation Balance with heating and cooling rates of 12 °C/min, while applying a steady field of 200 mT. The common feature of thermomagnetic experiments is that cooling curve is far above heating curve up to the temperature of 700 °C/600 °C (Fig. 3a), suggesting that magnetic mineral with high magnetization has been formed during heating. And the main decreasing of magnetization occurred at ~300 °C except for a small kink at ~400 °C. In order to better define temperatures at which magnetochemical changes occurred, J – T curves at a series of increasing temperatures have been determined (Fig. 3b). It is obvious that heating and cooling curves are basically reversible if the heating temperature is lower than 300 °C. The heating and cooling curves are characterized by irreversible behavior up to 400 °C, suggesting that magnetochemical changes occurred at the heating temperature intervals between 300 and 400 °C. A distinct Curie temperature is determined as the

Fig. 2. $^{40}\text{Ar}/^{39}\text{Ar}$ age spectrum plots for sample JG-1 (a) and JG-5 (b).

intersection of the descending curve with the tangent to the subsequent paramagnetic tail, $\sim 300^\circ\text{C}$, indicative of Ti-rich titanomagnetite in the studied sample. Such thermomagnetic behavior suggests that obvious alteration of the magnetic mineralogy had occurred if the heating temperature is higher than 300°C . It might thus be deduced that no reliable palaeointensity estimates could be obtained above 300°C .

Magnetic hysteresis loop determinations (Fig. 4a) were conducted on small chips taken three to four samples from each lava using a Micromag 2900 al-

ternating gradient force magnetometer. After removal of a small paramagnetic contribution (given by the slope at high fields) values of parameters H_{cr} , H_{r} , M_{s} , and M_{rs} were obtained for all measured samples. When plotted on a Day diagram (Day et al., 1977), the results (Fig. 4b) indicate that the magnetic carriers in all samples fall within the pseudo-single-domain (PSD) size range or a mixture of single domain and multidomain grains (Gogutchiaichvili et al., 2001a). It needs to be kept in mind, however, that reliability of the domain-state estimation using room temperature

Fig. 3. (a) Representative results of thermomagnetic curves (J - T) of samples from each lava flow. (b) Progressive J - T curves for lava #4.

hysteresis parameters in terms of the plot of magnetization ratio versus coercivity ratio has been questioned (Gogutchichvili et al., 2001b).

The texture and composition of magnetic minerals were studied with electron microprobe analysis through polished thin sections by using a CAMECA-CAMEBAX SX51 at IGG-CAS. The

basalt shows a typical and original intergranular texture (Fig. 5), almost no chemical alteration caused by the fluid or weathering could be found in plagioclase, pyroxene, and magnetic minerals. The average chemical compositions of 10 magnetic minerals are TiO_2 of 21.02 wt.%, TFeO of 64.39 wt.%, Al_2O_3 of 3.49 wt.%, and MgO of 4.34 wt.%, others less than

Fig. 4. (a) Example of hysteresis loop with maximum field of 1.5 T, but this figure only show the part below 300 mT. (b) Day plot of hysteresis parameters for samples from each lava flow.

Fig. 5. Backscatter electronic (BSE) image shows a typical and original intergranular texture in lava #3, with no alteration caused by fluid or weathering could be found.

1.00 Wt.%, strongly suggesting that these magnetic minerals are titanomagnetite. No significant solid solutions like ilmeno-hematites/titanomaghemites were observed.

3.2. Thellier's experiments

Before carrying out palaeointensity experiments, it is important to determine whether the titanomagnetite

Table 2
Palaeofield results

Lava #	<i>N</i>	Dec	Inc	α_{95}	Samples	T_s-T_e	<i>n</i>	<i>f</i>	Gap	<i>q</i>	$F_a \pm \text{S.D.}$	$F \pm \text{S.D.}$	VDM* $\pm \text{S.D.}$
1 ^a	9	24.4	63.6	5.2	JG1-2	100–260	5	0.52	0.73	4.66	26.09 ± 2.14	30.43 ± 5.16	5.44 ± 0.85
					JG1-8	100–260	5	0.57	0.66	13.35	34.77 ± 0.97		
2 ^a	8	31.9	65.2	3.4	JG2-1	100–285	6	0.57	0.73	6.61	29.08 ± 1.81	24.8 ± 4.15	3.27 ± 0.67
					JG2-5	150–285	5	0.49	0.68	4.78	25.47 ± 1.76		
					JG2-9	100–285	6	0.79	0.79	9.05	19.87 ± 1.38		
3 ^a	9	29.5	61.8	3.0	JG3-4	100–260	5	0.64	0.73	12.33	18.92 ± 0.72	23.27 ± 1.92	3.89 ± 0.32
					JG3-1	150–285	5	0.67	0.74	8.91	24.73 ± 1.38		
					JG3-2	200–285	4	0.55	0.60	10.82	26.17 ± 0.8		
4 ^a	11	23.4	64.3	4.2	JG4-1	100–285	6	0.54	0.73	8.97	25.67 ± 1.13	22.27 ± 1.89	3.43 ± 0.31
					JG4-2	100–260	5	0.64	0.75	28.76	20.82 ± 0.35		
					JG4-8	150–285	5	0.56	0.71	1.57	20.31 ± 5.16		
1–4	37	27	63.7	1.9								24.72 ± 1.24	3.83 ± 0.2
5	13	16.3	56.7	7.2	JG5-1	100–285	6	0.66	0.7	21.42	26.54 ± 0.57	24.33 ± 1.15	4.36 ± 0.21
					JG5-2	100–285	6	0.85	0.78	16.68	23.59 ± 0.94		
					JG5-7	150–285	5	0.71	0.72	25.73	22.85 ± 0.45		

Note: *N*, the number of samples used for palaeodirection determination; *n*, the number of NRM points used for palaeointensity determination; Dec and Inc, declination and inclination individually; α_{95} , radius of circle of 95% confidence about the direction; T_s/T_e , starting/ending temperature used for palaeointensity determination; *f*, the percentage of NRM used for palaeointensity determination; gap, an index of how well the NRM–TRM points are distributed along a straight line segment; *q*, quality factor defined by function $|b|/g/\sigma(b)$ where *b* is the slope and $\sigma(b)$ is its standard deviation (Coe et al., 1978); F_a and S.D., palaeointensity estimate for individual samples and standard deviation; *F*, mean palaeointensity for each lava; VDM*, virtual dipole moment.

^a Lavas are palaeomagnetically indistinguishable.

is “primary” or weathering-derived, as secondarily weathered titanomagnetite may carry a chemical remanent magnetization and invalidate Thellier palaeointensity experiments (Riisager et al., 2002). As above mentioned, electron microprobe observations on polished thin sections indicated that the minerals including Fe-minerals and plagioclase did not undergo metasomatic reactions or inversions after these rocks formed. It is evident that titanomagnetite was likely to be primarily formed during magma cooling from the melt.

Thellier’s palaeointensity experiments were conducted on at least 10 samples from each lava flow in argon atmosphere following Coe’s methodology (Coe, 1967), with a 40 μ T laboratory field applied during both the heating and cooling cycles. Sample selection was consistent with the criteria of Zhu et al. (2001). Temperature steps, chosen with reference to thermomagnetic curves, were lower than 300 °C. In order to constrain and detect possible magnetomineralogical changes occurring during heating, systematic partial thermoremanent magnetization (pTRM)-checks were performed after each pTRM acquisition.

A total of 61 samples from five lavas were selected for determining palaeointensity. Four representative NRM–TRM plots are shown in Fig. 6. The quality of the palaeointensity determinations was evaluated using the statistical parameters developed by Coe et al. (1978) and shown in Table 2. The NRM fraction (f) is the percentage of the NRM taken into account for the palaeointensity determination. We consider only linear segments defined by at least four points and that span at least 40% of the total extrapolated NRM (that is f values larger than 0.4) (e.g. Coe et al., 1978; Prevot et al., 1985) and the pTRM remains positive. The gap factor (g) is a measure of how the NRM–TRM points are distributed along the linear segment. The more irregular the spacing between the selected points of the segment, the more the gap factor will decrease from unity. The assessment of quality (q) is a function of these two parameters.

We found that 14 samples from five lavas met the criteria discussed above and yielded reliable palaeointensity estimates. Palaeointensities were calculated using a least squares fit to a straight line through the selected temperature interval. All samples behaved in a linear manner showed good internal consistency with low within-flow scatter. The mean

Fig. 6. Representative NRM–TRM plots (left) and associated orthogonal diagrams (right) for Jianguo samples. In the NRM–TRM plots, filled (open) circles denote data points accepted (rejected) from the best-fit line used to define the Palaeointensity values, solid triangles are pTRM-checks. In the orthogonal diagrams, the solid (open) symbols refer to the horizontal (vertical) plane. All studied samples were subjected to a hybrid demagnetization technique, in which thermal demagnetization up to 280 °C is followed by AF demagnetization up to 30 mT. The numbers refer to temperatures in °C.

Fig. 7. Examples of rejected NRM–TRM diagrams labeled by open circles with temperatures in °C shown to upper right; triangles are pTRM-checks.

palaeointensity value calculated for each flow ranges from 22.27 to 30.43 μT (Table 2), and the corresponding virtual dipole moment (VDM*) values ranges from 3.61×10^{22} to $4.99 \times 10^{22} \text{ Am}^2$.

Typical non-ideal diagrams are shown in Fig. 7. For examples, samples JG1–4 and JG5–5 exhibited a complex NRM–TRM diagram with negative pTRM-checks and no alignment of the successive points. No palaeointensity estimations can be obtained from those samples. The reason for failed experiments is probably magnetochemical alterations of titanomagnetite occurred during heating at lower temperatures.

4. Discussion and conclusion

It is noted that a stable characteristic remanent magnetization was obtained from the majority of samples (50 out of 64) after demagnetization to a peak temperature of only 100 °C though no reliable palaeointensity data could be obtained. Lava-mean directions and statistical parameters were calculated for each flow assuming a Fisherian distribution (Table 2). The results show that all five flows possess a normal polarity.

It is worthy to clarify whether or not five studied lava flows identified in field are palaeomagnetically independent. A *F*-test (McFadden and Lowes, 1981) to compare the mean directions of adjacent individual flows indicated that lavas #1–4 were indistinguishable. These palaeomagnetically related lava flows may imply either subsequent eruption in a relatively short time interval and/or flow number overestimation in field. Thus, these four lavas were combined as lava #1–4

in further analysis. The calculated mean direction of lava #1–4 is $D = 27^\circ / I = 63.7^\circ$ ($\alpha_{95} = 1.9^\circ$). The mean direction, calculated from all accepted samples because of only two time-independent lava flows, are $D = 23.8^\circ / I = 62^\circ$ ($\alpha_{95} = 2.4^\circ$) and its corresponding the mean VDM* is $(4.07 \pm 0.2) \times 10^{22} \text{ Am}^2$. It should be kept in mind, however, that the averaged paleofield values obtained are not necessarily free of effects caused by an insufficiently averaged paleosecular variation because only two time-independent lava flows were studied. On the other hand, the mean declination determined in this study is lying $\sim 17^\circ$ away from that derived from Eurasia apparent polar wander paths (Besse and Courtillot, 2002). This indicates that the studied area is not necessary free of regional deformation (Bureau of Geology and Mineral Resources of the Liaoning Province, 1989; Zhu et al., 2003b). But no significant latitudinal placement was revealed within 95% confidence. Thus, it is reasonable to calculate the VDM*.

Our study, employing Thellier–Thellier absolute palaeointensity method, provides a reliable determination of geomagnetic field strength for the interval of 105.5–102.2 Ma. We now attempt an analysis of palaeofield strength within the CNS, although there are relatively few palaeointensity estimates involved. In order for a given published set of determinations to be considered in our analysis the data are selected from IAGA palaeointensity database Pint2003 (Perrin et al., 1998; <ftp://dstu.univ-montp2.fr/pub/paleointdb/>). Indeed, there are no criteria presently available that could provide an assurance in the absence of observatory geomagnetic field measurements. Therefore, in order to analyze the palaeointensity database, we temporally used the pTRM-check criterion. Needless to say, a palaeointensity estimate that passes the pTRM-check is certainly not guaranteed to be accurate either (Thomas and Biggin, 2003). The following datasets within the CNS satisfy the selecting criteria: whole rock (Sherwood et al., 1993; Tanaka and Kono, 2002); SBG study of the Hole 543A (Pick and Tauxe, 1993; Juarez et al., 1998; Selkin and Tauxe, 2000); and single plagioclase crystals (Tarduno et al., 2001, 2002).

When combined these selected data with our palaeointensity result, a mean VDM of $(5.17 \pm 0.8) \times 10^{22} \text{ Am}^2$ is obtained (Fig. 8) within the CNS, a value that is approximately half of that of the present-day

Fig. 8. Collections of VDM/VADM data for the entire CNS (lightly shaded part). Solid (open) circles represent data from whole rock (SBG), triangle represents data from single plagioclase crystals. Solid star stands for the whole rock result of this study.

field. If we now remove the data obtained from SBG, it then gives a mean value of $(5.37 \pm 0.82) \times 10^{22} \text{ Am}^2$, which is very similar to $(5.17 \pm 0.8) \times 10^{22} \text{ Am}^2$. Hence, SBG data have been included though reliability of SBG study is now questioned (Heller et al., 2002).

The strength of the Earth magnetic field within the CNS is an important subject, not only to discuss the relationship between polarity reversal frequency and palaeointensity, but also to consider the geodynamo within the Earth. It is, therefore, valuable to compare mean palaeointensity within the CNS with that out of it, i.e., the entire 65.7 Ma interval (Shi et al., 2002). The mean dipole moment of $(5.17 \pm 0.8) \times 10^{22} \text{ Am}^2$ within the CNS is close to the mean value of $(5.26 \pm 2.89) \times 10^{22} \text{ Am}^2$ for the entire 65.7 Ma interval (Shi et al., 2002), suggesting that the mean strength of the Earth magnetic field within the CNS is comparable to that out of the CNS though some high paleointensities using the plagioclase crystals have been reported (Tarduno et al., 2001, 2002). Further, we noted that the paleointensity values within the CNS exhibit a very highly variable nature (Fig. 8). It may be associated with the dynamo processes during the CNS, if those palaeointensity estimations are correct. Clearly, more data are needed to better understand relationship between this long-term variation of palaeointensities and geodynamo process.

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