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Paleointensity of the Geomagnetic Field in the Last Half-Million Years in Regions of the Red Sea and South of the Mid-Atlantic Ridge

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Abstract—The paleointensity of a geomagnetic field is determined on the basis of the standard Thellier's method using the remanent magnetization (H_a) of basalts of the rift zone of the Red Sea (with an age of 0.5 million years) and south of the Mid-Atlantic Ridge (with an age of 0.1–0.3 million years). Samples were selected whose natural remanent magnetization (NRM) was mainly one-component and the measured Curie temperatures of the titanium–magnetite fraction (T_C = 205–250°C) were close to the calculated values. Samples with T_C = 580°C were also used. It was shown that NRM of basalts with both high and low T_C values have a thermoremanent nature. The studies have revealed that the law of independence and additivity of partial thermoremanent magnetizations (PTRMs) is satisfied for the collected samples with an error of no more than 6% and the error of determination of the field of TRM creation using the Arai–Nagata diagram is no more than 10%. The value H_a = 32 A/m determined from NRM of basalts south of the Mid-Atlantic Ridge appears to be approximately equal to the contemporary intensity of the geomagnetic field (H_c) in this region. In the Red Sea region, H_a = 77.5 A/m, which is 2.5 times higher than the current value. It is suggested that during formation of the basalts of the Red Sea rift zone the geomagnetic pole was located in that region and the virtual dipole moment was 35% higher than its contemporary value.

Keywords: geomagnetic field, paleointensity, basalts, Thellier method.
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INTRODUCTION

It is known that the Earth's magnetic moment and magnetic field intensity change with time. In the last 450 years, the value of the dipole magnetic moment of the Earth decreased by almost 20% [11].

The main geomagnetic field is defined as the field obtained by averaging its measured values over the time interval exceeding 1 year and over an area of more than 10^6 km^2. The international standard for the main geomagnetic field is the International Geomagnetic Reference Field (IGRF) [4]. However, it has been shown that in the late 15th century the field of a dipole located in the Earth's center and aligned along the axis of rotation was a good approximation of the geomagnetic field. According to the Gauss theory of geomagnetic fields [11], the first three terms in the Gauss infinite series describe the field of a dipole inclined to the Earth's rotation axis. The geomagnetic dipole determined from measurements of the magnetic field intensity on the Earth's surface at one point is named a virtual dipole and the appropriate geomagnetic poles are called as virtual poles. These notions are widely used in paleomagnetism. In studying the paleomagnetism of mountain rocks, we may estimate the values of the virtual dipole moment (VDM) in past geological epochs. Numerous data on VDM determination known from the literature testify to its substantial variations in time [7, 9].

In order to study the VDM distribution in time, the international IAGA paleointensity database was created recently on the website of the National Geophysical Data Center in Boulder (Colorado) [15]. This database (with the use of additional information from the database of the Borok observatory [12]) is supplemented with the data of studies [16, 17] and it includes 3203 VDM values. The results of VDM determination over the last 2000 years and for 150000 years are presented in Fig. 1. From these data it is seen that the results of VDM determination even for relatively young rocks may differ twofold from their contemporary value and the average VDM value, e.g., 2500 years ago, was 1.5 times greater than the contemporary value.

This work is aimed at determination of the ancient magnetic field (H_a) from the NRM values (I_n) of basalts from rift zones of the Red Sea and the southern part of the Mid-Atlantic Ridge (MAR).

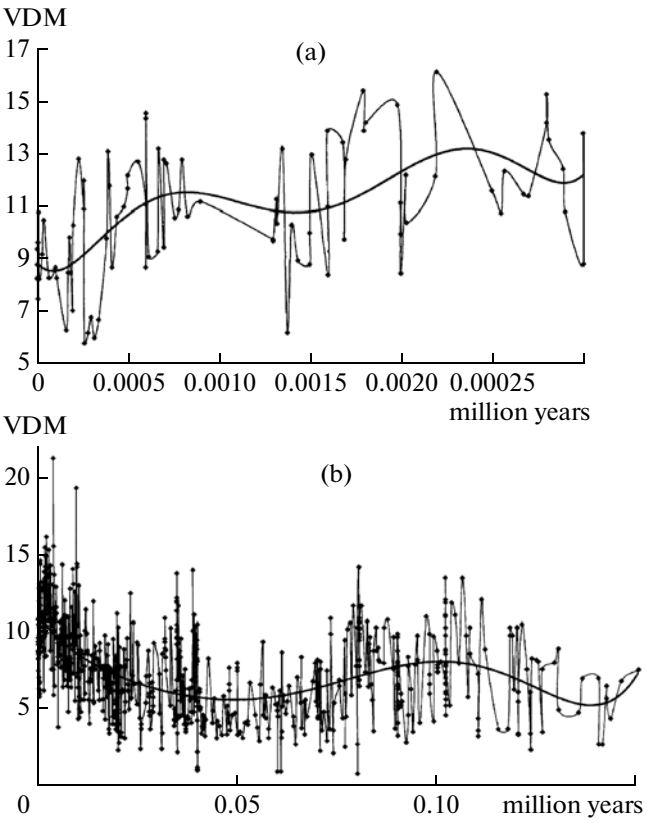

Fig. 1. The Earth's VDM variation in time according to data of the National Geophysical Data Center in Boulder (Colorado) over (a) the latest 2000 years and (b) 150 000 years.

OBJECTS OF EXAMINATION

For determination of the intensity of the ancient geomagnetic field, basalt samples (53-4, 57, 62-3, 65-1, 66-4, 68-2, 71-3, 72-5) were examined that were taken from the Red Sea bottom with the *Pisces* manned submersible within the expedition (1980) of the Shirshov Institute of Oceanology of the Academy of Sciences of the Soviet Union, in the rift zone area with coordinates of 17°56′–18° N and 40°04′–40°06′ E. The age of the investigated basalts of the Red Sea according to [8] is under 0.5 million years. The information concerning the magnetism of the Red Sea rift zone is given in [5].

We also examined basalt samples from the MAR south (sample 51-3, the coordinates of the selection point are 54°40′ S, 1°00′ W) and from the Bouvet ridge (samples 30-18, 31-37, coordinates of the selection points are 54°00′ S, 3°30′ E) which were obtained by the expedition onboard the r/v *Akademik Nikolaj Strakhov* (18th cruise) in 1994. The magnetism of basalts from the MAR south and the Bouvet ridge was considered at some length in [6]; the basalt ages are 0.1 to 0.3 million years (Table 1).

The magnetic field intensity, calculated using a IGRF/DGRF model [18] of geomagnetic field for 2000, is 28.4 A/m in the Bouvet ridge area and H_e = 30.7 A/m in the region of the Red Sea (40°06′ E, 17°56′ N) where our samples were taken.

INSTRUMENTS AND METHODS OF EXPERIMENTS

The magnitude and direction of the remanent magnetization of samples with respect to the chosen coordinate system were measured at room temperature using a JR magnetometer. The value of the magnetic susceptibility κ_0 was measured using a KLY-2 instrument. The Curie temperature was determined from the temperature dependence of the magnetic susceptibility $\kappa_0(T)$ using the MFK1-A instrument with a CS-3 thermal attachment and also from the dependence of the magnetization $I(T)$ in the field H = 0.24 T using a vibration magnetometer. Temperatures of blocking were determined from the destruction curve of $I_n(T)$ and $I_{rs}(T)$.

Hysteresis and thermomagnetic properties, as well as normal magnetization curves, were investigated using a vibration magnetometer VMA-1 [2]. The coercive (H_c) and residual coercive (H_{cr}) forces were determined by the "thickness" of the hysteresis loop in demagnetization once the sample was magnetized in the 0.4-T field. Using the micro weber meter by the Weiss and Forrer method, we measured the saturation magnetization I_s. For estimation of the contribution of the paramagnetic part to the I_s determination, the $I(H)$ dependence was taken with several field values and the correction for the paramagnetic constituent was applied by subtracting the linear component of $I(H)$ from the total signal.

Using a VMA-1 vibration thermomagnetometer we also obtained destruction curves of the residual saturation magnetization during the heating in the space bounded by a two-layer permalloy screen.

RESULTS

Component Composition and Stability of I_n(NRM)

As is known, one of the stages of a paleomagnetic study is the estimation of the component composition of the NRM I_n, which is performed by checking the direction of the remanent magnetization vector during the I_n thermal demagnetization (the Zuiderveld vector, Fig. 2). Among the Atlantic basalts, we chose the samples whose I_n mainly had one component (51-3 and 31-7 in Fig. 2a) or the second component contributed slightly to I_n and was demagnetized already after heating to T = 100°C (sample 30-18 in Fig. 2b). Some samples of the Red Sea basalts also had a one-component I_n (72-5, 53-4, 65-2, and 57 in Figs. 2c, 2d, and 2f). The other groups of samples (65-1 and 71-3 in Figs. 2d and 2e) were previously subjected to thermal cleaning. Here it appeared that the maximum temperature of heating, after which the Zuiderveld vector stopped turning, varied for different samples from 100

Table 1. Magnetic properties of basalts of the bottom south of the Atlantic Ocean and of the Red Sea bottom

Sample	I_n, A/m	$\kappa_0 \times 10^{-2}$, SI units	Q_n	I_{rs}/I_s	H_c, mT	H_{cr}/H_c	T_c, °C	$T(50\%I_n)$, °C	T_{bmax}, °C	H_a, A/m	Age, million years
30-18(2)[1]	4.6	1.3	12.5		–	–		160	174	22	0.1
31-37(2)[1]	17.8	0.89	70		14.5	1.38		175	270	32	0.2
51-3(1)[2]	28.9	0.98	104		29.3	1.29		200	291	33	0.3
53-4[3]	47	0.39	392		56	1.34		347	470		Under 0.5
57[3]	39.9	0.4	325		44	1.27		346	477	89	
62-3(5)[3]	66.5	0.45	481		75	1.67		375	360		
65-1(3)[3]	53.6	0.22	794		45.2	1.54		345	570	76	
66-4(4)[3]	107.8	1.32	266	0.35–0.5	24	1.23	280 470 580	280	362		
68-2(6)[3]	70.3	1.22	188		20	1.25		260	361		
71-3(5)[3]	52	1.61	105	0.35	16	1.25	250	250	350	79	
72-5(5)[3]	64.7	1.9	111	0.29	14	1.28	205	240	320		
65-3(1)[3]	23	0.38	223	0.22	34	1.35					
65-3(8)[3]	55.7	0.64	318	0.28							
65-2[3]	9.1	1.58	213	0.19	24.7	2.1	480 580	420	540	77.5	

Notes: [1] Basalts of the Bouvet ridge.
[2] Basalts of the MAR south.
[3] Basalts of the Red Sea.

to 300°C. Samples with a large turn of the Zuiderveld vector during the demagnetization were rejected. The I_n stability in thermal demagnetization differed substantially for various samples (Fig. 3). The comparison of data presented in Figs. 2 and 3 shows that on completion of the Zuiderveld vector turning in demagnetization of samples the value of the remanent magnetization decreased by no more than 5%.

The temperature of demagnetization (to 50% of I_n) for samples of the Red Sea basalts (see Fig. 3 and Table 1) varied approximately from 240 to 420°C. Among the examined collection of basalts of the Red Sea, samples whose I_n was destroyed only after heating to the temperature equal to the Curie temperature of the magnetite were found (580°C). for example, I_n of samples 65-2 and 65-1 in heating to 550°C was destroyed by only 95% and the I_n of sample 65-1, by 80%.

Comparison of the magnetic characteristics of the investigated basalts of the Atlantic and the Red Sea demonstrates the following. The maximal temperature of blocking (T_{bmax}) of the Red Sea basalts determined from the destruction of the remanent saturation magnetization I_{rs} during the heating of the samples varied from 320 to 570°C. The examined samples had fairly high values of NRM (I_n = 39.9–107.8 A/m) and the Koenigsberger factor (Q_n = 105–794).

The basalt samples of the MAR south had smaller values of the respective quantities (I_n = 4.6–28.9 A/m, Q_n = 12.5–109, and T_b < 290°C) than basalts of the Red Sea. They are also characterized by lower I_n stability in the thermal demagnetization of the samples. The temperature of demagnetization to 50% of I_n varied approximately within 160–200°C; to 90% of I_n, within 250–300°C. The maximal blocking temperature determined by I_{rs} varied from 174 to 291°C (see Table 1).

The Phase Composition of Ferrimagnetic Grains

The I_n stability during thermal demagnetization is substantially governed by the Curie temperature of ferrimagnetic phases, NRM carriers. We measured the Curie temperature of ferrimagnetic phases using curves of temperature dependence of the magnetic susceptibility $\kappa(T)$. As is seen from Fig. 4, the dominant phase with the low Curie temperature (T_C = 225–250°C) could be detected rather confidently for some samples (72-5 and 71-3). Two phases with T_{C1} = 480°C and T_{C2} = 580°C were found for sample 65. By curves $\kappa(T)$, even three phases with T_{C1} = 280°C, T_{C2} = 470°C, and T_{C3} = 580°C could be revealed on sample 66-4.

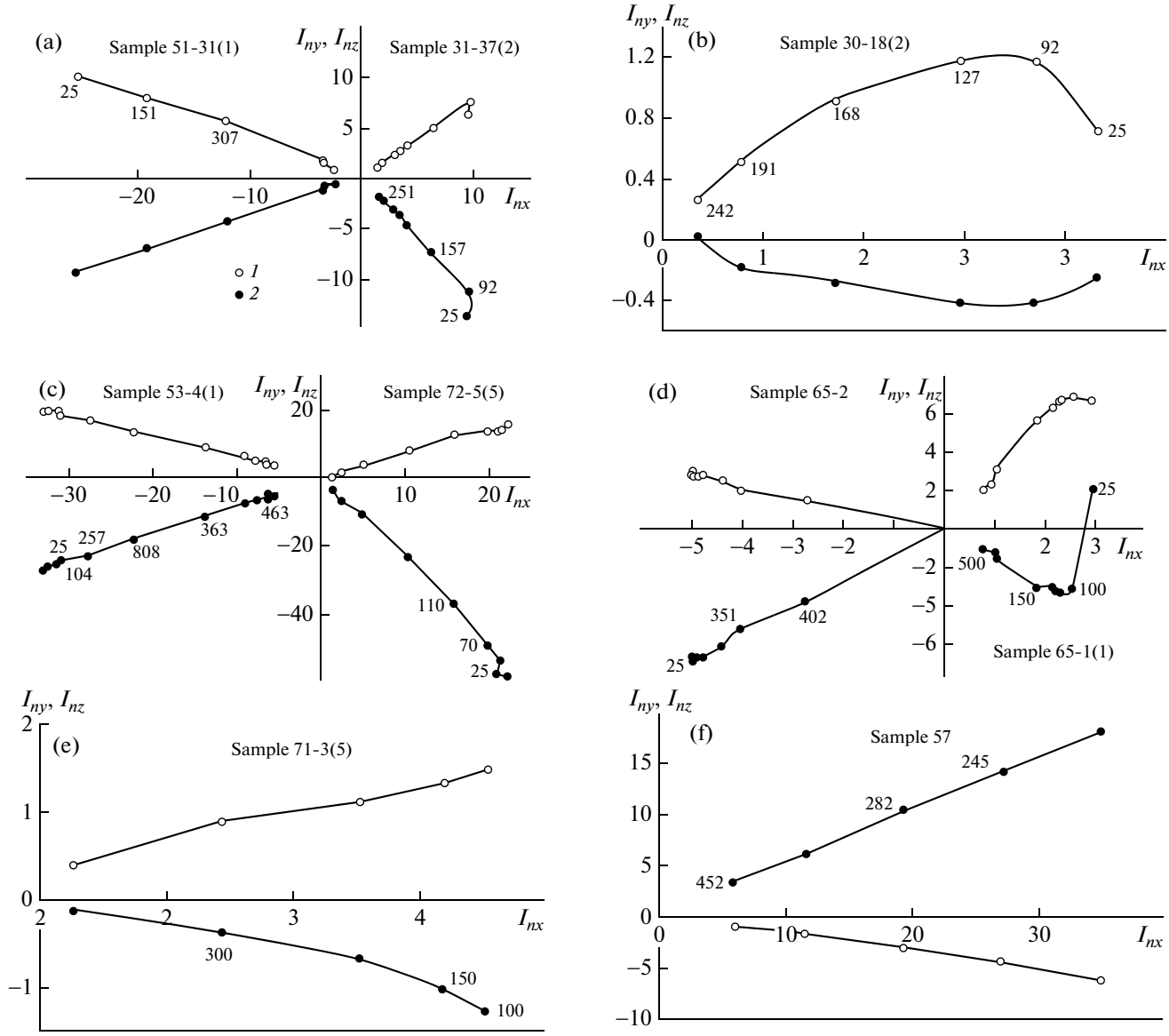

Fig. 2. Zuiderveld diagrams obtained during the thermal demagnetization of samples from the natural state. The numerals near points indicate the temperature the sample was heated to in the cycle. (a) Basalt samples from the Bouvet ridge. On the left, sample 51-3(1); on the right, sample 31-37(2). (b) Basalt sample 30-18(2) from the MAR south. (c) Basalt samples from the Red Sea. On the left, sample 53-4; on the right, sample 72-5(5). (d) Basalt samples from the Red Sea. On the left, sample 65-2; on the right, sample 65-1(1). (e, f) The Red Sea basalt samples 71-3(5) and 57. (*1*) Projection of magnetization onto the *xy* plane. (*2*) Projection of magnetization onto the *xz* plane.

The comparison of T_C and $T_{50\%\,In}$ values (see Table 1) shows that in the majority of cases once samples are heated to the Curie temperature of the low-temperature magnetic phase, no more than 50% of the NRM is demagnetized, i.e., approximately 50% of I_n is associated with the low-temperature phase. We may suggest that this part of the magnetization is primary. An electron probe analysis of ferrimagnetic grains was performed on individual samples with both low (samples 71-1 and 72-5) and high (samples 66-4 and 65-3) Curie temperatures. According to these data, the ferrimagnetic fraction of the Red Sea basalts is represented by spinellides, whose grains have a dendrite structure (Fig. 5). The TiO_2 content in the samples varied from 11.63 to 17.66% (Table 2). Low concentrations of MgO (0.43%) and MnO (0.46%) were found in individual samples. The ulvöspinel concentration (x), on assumption that grains consist of titanium magnetites with the stochiometric composition $x\mathrm{Fe_2TiO_4}\ (1-x)\mathrm{Fe_3O_4}$, varied from $x = 0.34$ to $x = 0.55$. The Curie temperature T_C^{calc} calculated using the dependence $T_C = f(x)$ [3] varied from 205 to 330°C (see Table 2).

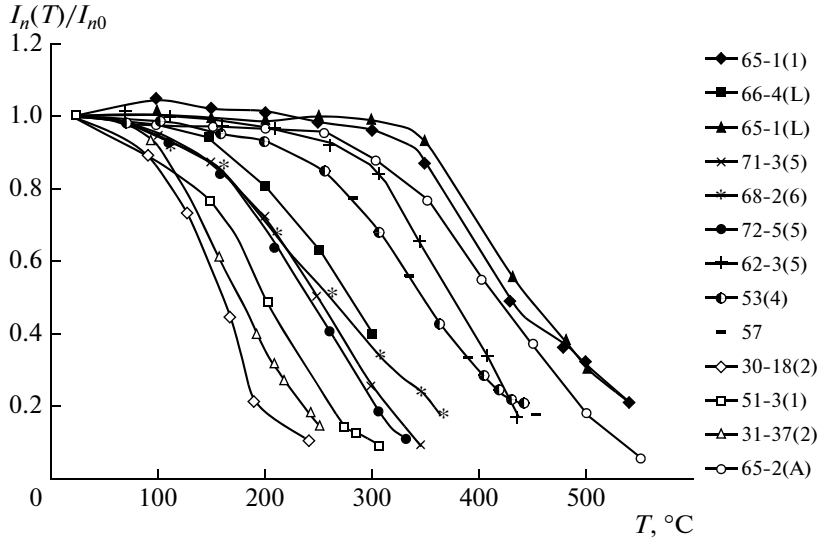

Fig. 3. Curves of I_n destruction in per-unit during thermal demagnetization of basalt samples from the Red Sea and the southern Atlantic.

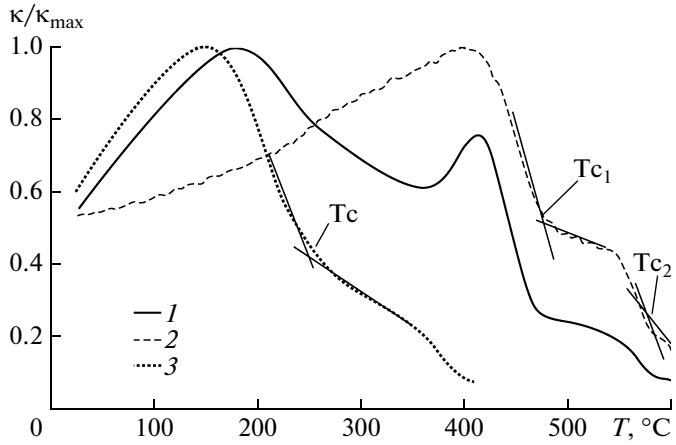

Fig.4. The temperature dependences of k/k_{max} for samples (*1*) 66-4, (*2*) 65-2, and (*3*) 71-3(5) of the Red Sea basalts.

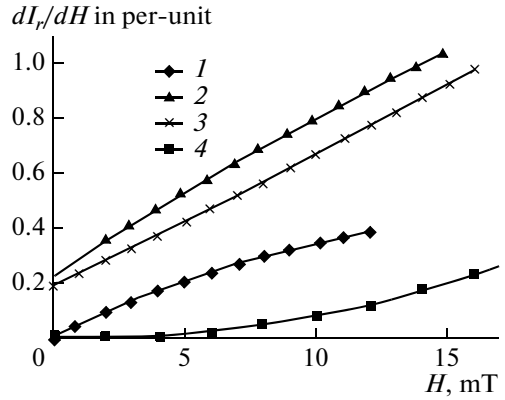

Fig. 5. Coercive spectra of normal magnetization of sample 65-1(2) from different states: (*1*) from natural state with I_n, (*2*) from AZS, (*3*) from state I_{rT}, and (*4*) from ZS.

The Curie temperature T_C measured on samples 71-3 and 72-5 appeared close to T_C^{calc} (see Tables 1 and 2). It is obvious that the larger part of I_n is primary on the samples with Curie temperatures that are not very high.

According to the data of thermomagnetic analysis, the ferrimagnetic phases with Curie temperatures of 480°C and 580°C are responsible for the magnetic properties of 65-1 and 65-2 basalts. However microprobe analysis demonstrated that there is a great quantity of TiO_2 (11.63% to 14.76%) in ferrimagnetic grains. Titanium magnetite of stochiometric composition with this TiO_2 concentration must have Curie temperatures of $T_C = 275–330$°C [3]. The magnetite's type of Curie temperature when there is a large concentration of TiO_2 testifies to the fact that the initial titanium magnetite in these basalts was subjected to heterophase oxidation. The absence of a Zuiderveld vector turn on these samples at high temperatures is evidence of the coincidence of the directions of remanent magnetizations formed on the magnetic phases with different Curie temperatures.

If phases with high Curie temperatures were produced at the stage of basalt formation, i.e., at $T >$ 600°C, then the natural magnetization of the magnetite phase will have the thermal remanent nature and the magnetic properties should correspond to the high-temperature state. If the high-temperature phase was created at $T < T_C\,(T_b)$, then I_n must be of a chemical nature and magnetic properties should conform to the low-temperature state. To prove the validity of either statement, we compared the coercive spectra of

Table 2. Geochemical characteristics of grains of spinellides included in the composition of basalts of the Red Sea rift

Sample	MgO	Al_2O_3	SiO_2	CaO	TiO_2	Cr_2O_3	MnO	FeO	ZnO	Total	x*	T_{calc}, °C
	mass %											
66-4L	0.25	3.36	4.04	0.59	16.47	0.00	0.00	69.89	0.00	94.60	0.494	210
66-4L	0.26	2.61	2.24	0.27	17.66	0.00	0.39	71.04	0.66	95.15	0.514	205
71-3(4)	0.00	2.43	0.65	0.28	17.06	0.00	0.00	73.26	0.00	93.69	0.500	210
71-3(4)	0.43	3.40	5.69	0.65	14.06	0.00	0.46	69.12	0.39	94.19	0.424	275
71-3(4)	0.35	2.43	0.79	0.32	19.29	0.00	0.44	70.92	0.53	95.07	0.554	180
65-3(1)	0.00	2.74	3.05	0.50	13.31	0.00	0.00	75.70	0.00	95.30	0.393	295
65-3(1)	0.26	2.50	1.91	0.43	11.63	0.00	0.00	76.94	0.67	94.34	0.343	330
65-3(1)	0.43	2.98	7.54	1.27	12.23	0.00	0.00	72.08	0.00	96.54	0.375	300
65-3(8)	0.00	2.45	1.90	0.70	14.76	0.00	0.36	74.31	0.00	94.49	0.429	275
65-3(8)	0.00	2.56	2.29	0.49	13.17	0.00	0.41	76.68	0.42	96.02	0.377	300
65-3(8)	0.22	2.35	2.56	0.43	14.39	0.00	0.00	74.14	0.53	94.64	0.428	275
72-5	0.00	2.45	1.80	0.35	16.43	0.62	0.62	78.36	0.00	100.0	0.491	210

* Fraction of ulvöospinel in solid solution of titanium magnetite.

normal magnetization [1] obtained from different states (see Fig. 5). The coercive spectra of normal magnetization contained from the absolute zero state (AZS) and from the I_{rT} state are almost completely identical. The spectrum of the normal magnetization from the zero state (ZS) corresponding to the low-temperature origin is substantially different than the spectra obtained from AZS and the I_{rT} state, which correspond to a high-temperature origin. The normal magnetization spectra obtained from the natural state lie higher than spectra obtained from ZS and they are closer to the spectra of normal magnetization from the I_{rT} state. Such a pattern was observed for samples with both low and high Curie temperatures T_C (T_b). On this basis, we may state that the part of the Red Sea basalts' NRM whose carrier is a phase that is close to that of magnetite, also has a thermal remanent nature and is primary. We previously suggested [5] that the magnetite phase formation in some samples of the Red Sea basalts occurred at the stage of their formation.

Determination of the Field of NRM Formation in the Basalts of the Red Sea and MAR South Using the Thellier Method

We used the Thellier method [18] to determine the field of NRM formation in basalts.

The criterion for the applicability of this method is that the law of additivity and independence of partial TRMs is satisfied for the investigated rocks; as is known [3], this law is valid only for single-domain grains. As was noted above, according to the electron-probe analyses, the ferrimagnetic fraction of the examined basalts is represented by grains of spinellides with sizes from 1 to 5 μm which had a dendritic structure. In spite of the rather large size of the ferrimagnetic grains, their magnetic state appeared to be close to that of single-domain grains. According to the criterion in [13], this is shown by the ratio values $H_{cr}/H_c = 1.23–2.08$ and $I_{rs}/I_s = 0.35–0.75$ (see Table 1). This is probably caused by a rather large contribution to the free energy of grains from the shape anisotropy because the dendrite length sometimes reached 40 μm.

Thermal magnetic studies demonstrated that the law of the independence of partial TRMs for the selected collection of samples was obeyed. for example, for sample 65-2 with the larger size of ferrimagnetic grains ($H_{cr}/H_c = 2.08$ and $I_{rs}/I_s = 0.35$) than that of the other samples, the sum of partial TRMs obtained in the temperature range of 20–550°C with a step of 50°C is only 6% larger than the TRM obtained in the same range.

The efficiency of the Thellier method for determination of geomagnetic field paleointensity was verified in the laboratory using the TRM created on samples 71-3(5) and 65-2. The experimental results are presented in Fig. 6 in the form of an Arai–Nagata plot [3, 10]. The magnetic field value determined from these data appeared to be 5–10% larger than the value of the field of TRM formation. This overestimated value of the field of TRM formation may be related to the presence of pseudo-single-domain grains in the ferrimagnetic fraction of the examined samples [10].

The application of the Thellier method for the determination of the field of formation of the primary remanent magnetization of basalt samples from the Red Sea and the MAR south showed that the dependence $I_i/I_{n0} = f(I_{rpT}/I_{n0})$ on the Arai–Nagata diagram may be approximated with sufficient accuracy by a linear function (see Fig. 7 and Fig. 8). For samples from

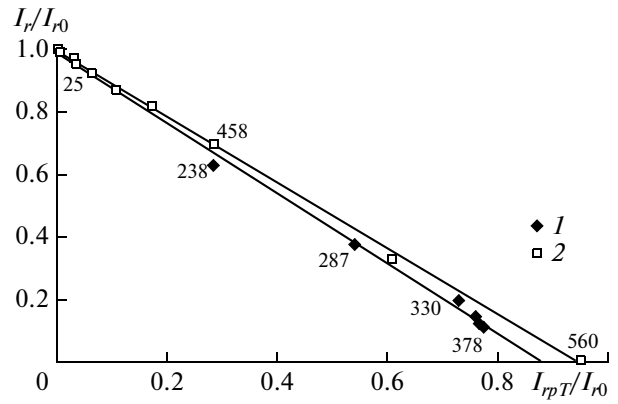

Fig. 6. Arai–Nagata diagram for samples (*1*) 71-3(5) and (*2*) 65-2. The numerals near the points are the maximum temperature of heating the sample in the cycle. Equations of straight lines: (*1*) $y = -1.11x + 14.37$, (*2*) $y = -1.08x + 0.98$.

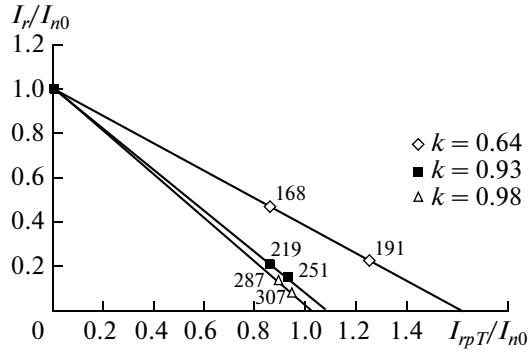

Fig. 7. Arai–Nagata diagram of the determination of the H_a of remanent magnetization formation for the MAR south basalts using the standard Thellier method (with $H_{lab} = 34$ A/m) for samples (*1*) 30-18(2), (*2*) 31-37(2), and (*3*) 51-3(1).

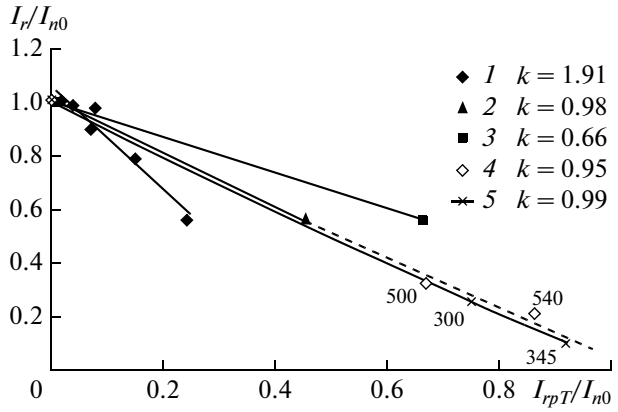

Fig. 8. Arai–Nagata diagram of the determination of H_a of NRM formation of the Red Sea rift zone basalts using the standard Thellier method for samples: 65-2 (with (*1*) $H_{lab} = 40$ A/m, (*2*) $H_{lab} = 80$ A/m, and (*3*) $H_{lab} = 120$ A/m); (*4*) 65-1 with $H_{lab} = 80$ A/m; (*5*) 71-3 with $H_{lab} = 80$ A/m.

the southern Atlantic, the value of the geomagnetic field paleointensity determined as $H_a = |k|H_{lab}$ (where k is the factor of linear approximation of data on the Arai–Nagata diagram (see Fig. 7), H_{lab} is the laboratory field) varied from 22 to 33 A/m (see Table 1). The certainty of the result of determination of $H_a = 22$ A/m on sample 30-18 is probably worse than that on samples 31-37 and 51-3 because in the first case a small turn of the I_n vector was observed during thermal demagnetization (see Fig. 1b). Thus, in the region of the Bouvet ridge 0.2–0.3 million years ago the value of the geomagnetic field paleointensity $H_a = 32$–33 A/m was fairly close to its contemporary value ($H_c = 28.4$ A/m).

For samples of the Red Sea basalts the k factor on the Arat–Nagata diagram in case of the I_{rpT} formation in the field $H_{lab} = 40$ A/m appeared to be more than unity (see Fig. 8); for $H_{lab} = 80$ A/m, it was close to unity. The value determined on three samples of Red Sea basalts was 76–79 A/m and on one sample the value 89 A/m was obtained, i.e., approximately 2.5 times greater than its contemporary value in that region ($H_c = 30.7$ A/m). Moreover, as seen from Table 1, the results of H_a determination on samples with both low (sample 71-3) and high (samples 65-1 and 65-2) degrees of oxidation of titanium magnetite coincide within the error limits. This coincidence also testifies in favor of our previous suggestion that the oxidation of the titanium magnetite of the basalt samples that were numbered 65, which were collected in the zone of nontransform displacement, occurred at the stage of their formation.

CONCLUSIONS

Based on the laboratory results, we can state that 0.2–0.3 million years ago the geomagnetic field paleointensity for the region of the southern Atlantic was only 10% larger than the contemporary geomagnetic field intensity in this region. In the region of the Red Sea, the paleointensity H_a 0.5 million years ago was 2.5 times larger than the contemporary geomagnetic field intensity in this region. We may suppose that during the formation of basalts in the rift zone of the Red Sea, the Earth's magnetic moment was substantially larger than at the present time if the geomagnetic latitude of this region remained unchanged. However, it is known that the position of the magnetic poles with respect to geographic poles changes, so we can put forward a second version of the explanation of our results: during the formation of the examined basalts, the geomagnetic pole was located in the Red Sea region and the geomagnetic equator was in the area of the Bouvet ridge. It is also possible that during the formation of the NRM of the Red Sea basalts, a fairly intensive geomagnetic anomaly of the East-Siberian type existed in this region.

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