PERGAMON

Available online at www.sciencedirect.com

SCIENCE @ DIRECT®

Physics and Chemistry of the Earth 28 (2003) 659–667

PHYSICS and CHEMISTRY of the EARTH

www.elsevier.com/locate/pce

Selected room temperature magnetic parameters as a function of mineralogy, concentration and grain size

C. Peters *, M.J. Dekkers

Palaeomagnetic Laboratory 'Fort Hoofddijk', Faculty of Earth Sciences, Utrecht University, Budapestlaan 17, 3584 CD Utrecht, The Netherlands

Abstract

A data set of room temperature magnetic parameters for several iron oxides and sulphides was compiled from the available literature. The aim was to propose the most effective methods for assessing mineralogy, concentration and domain state within environmental magnetic studies. Establishing the magnetic mineralogy is essential for a correct interpretation of concentration and grain-size indicative parameters. Hematite and in particular goethite are recognised by their high (remanent) coercivities. The ratio of saturation remanent magnetisation to susceptibility is suited for assessing mineralogy; for low coercivity minerals, high values of the ratio are indicative of pyrrhotite. Greigite and maghemite both have intermediate ratios, while very low values suggest the presence of (titano)magnetite. From the concentration-dependent parameters mass specific susceptibility, saturation remanent magnetisation and susceptibility of anhysteretic remanent magnetisation, susceptibility displayed the least grain-size dependence. Therefore, it is perhaps the best indicator of concentration, under the proviso of uniform mineralogy and when paramagnetic and diamagnetic contributions are taken into consideration. All minerals showed a decrease in coercivity and remanence ratios with increasing grain size for sizes larger than one micrometer. In contrast, the coercivity of goethite showed an increase with grain size. Assessment of domain state was complicated by very small and large grains displaying similar magnetic properties.

Keywords: Rock-magnetic parameters; Mineral magnetism; Grain size; Room temperature

1. Introduction

The classification of magnetic components within natural samples in terms of mineralogy, domain state, and concentration, is important for assessing the nature and origin of the components, e.g. post-depositional formation of bacterial magnetite or greigite within sediments. Many publications document the variation of magnetic parameters with grain size for individual minerals. Hunt et al. (1995), Peters (1995) and Peters and Thompson (1998) have summarised magnetic parameters for rocks and minerals. Here a detailed study of how magnetic parameters vary with mineralogy and grain size is presented which incorporates data that were not available in the mid-nineties. Individual minerals are considered from superparamagnetic to multidomain (MD) grain size. Data was compiled from a substantial number of publications with the aim of investigating the best ways to define mineralogy, concentration and domain state in order to maximise the information obtained from room temperature magnetic data of samples for environmental applications. Recent volumes that summarise the variability of magnetic properties and natural samples include (Dunlop and Özdemir, 1997) with most information pertaining to the titanomagnetite series (Maher and Thompson (1999); Walden et al., 1999). The latter two volumes review mineral magnetic methodology as applied to sediments and soils, for example to retrieve information concerning the paleoclimate record and/or the provenance area(s). Walden et al. (1999) present an overview of laboratory methods to acquire optimal magnetic information from a sample set. Natural samples are often characterised by a mixed magnetic mineralogy which complicates an environmental interpretation. In such cases, trends of individual minerals serve as basis for the interpretation. For the present contribution, a compilation was prepared of data of synthetic and natural samples of the iron oxides magnetite, titanomagnetite, maghemite and hematite, the iron oxyhydroxide goethite, and the iron sulphides pyrrhotite and greigite. Synthesis or upgrading

---

* Corresponding author. Fax: +31-30-253-1677.

E-mail addresses: peters@geo.uu.nl (C. Peters), dekkers@geo.uu.nl (M.J. Dekkers).

1474-7065/$ - see front matter © 2003 Published by Elsevier Ltd.
doi:10.1016/S1474-7065(03)00120-7

methodology, how grain-size fractions were obtained, and details of the instrumentation used can be found in the individual publications. During the compilation it was necessary to standardise units by converting some data from CGS units to SI units, and in other cases from volume specific to mass specific values. The conversion factors are listed in Appendix A.

2. Magnetic mineralogy

Combining magnetic parameters graphically is very useful for assessing magnetic mineralogy (e.g. Peters and Thompson, 1998). Fig. 1, the variation of the ratio of saturation remanent magnetisation to susceptibility, σ_{RS}/χ, versus the remanent acquisition coercivity, $(B_0)_{CR'}$, shows an example of the use of biplots for assesing mineralogy. Recent developments in the modeling of IRM acquisition curves into individual coercivity contributions estimates values of $(B_0)_{CR'}$ (Heslop et al., 2002). The outcome of coercivity component analysis from IRM acquisition curves is dependent on the initial state of the samples (quite often AF demagnetised, which yields higher coercivity values than when doing IRM acquisition starting from the NRM state of a sample). A more traditional form of the biplot shown in Fig. 1 is σ_{RS}/χ plotted against the remanent coercivity, $(B_0)_{CR}$, (cf. Thompson and Oldfield, 1986; Peters and Thompson, 1998). Both biplots display similar results; goethite and hematite are characterised by their high remanent (acquisition) coercivities. Goethite is distinguished from hematite by its higher values. For a given remanent (acquisition) coercivity pyrrhotite has higher ratios of σ_{RS}/χ than (titano)magnetite, maghemite and greigite. For the minerals exhibiting low remanent (acquisition) coercivities, only (titano)magnetite samples display σ_{RS}/χ values less than 4 kA m^{-1}.

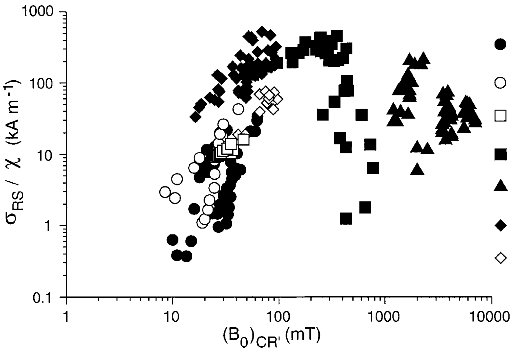

Fig. 1. Biplot of σ_{RS}/χ versus remanent acquisition coercivity for magnetite, titanomagnetite, maghemite, hematite, goethite, pyrrhotite and greigite. The data are from Bailey and Dunlop (1983), Clark (1984), Dankers (1978, 1981), de Boer and Dekkers (1996), Dekkers (1988, 1989), Dekkers and Linssen (1989), Dekkers and Schoonen (1996), Dunlop (1971), Hartstra (1982b), Maher (1988), Menyeh and O'Reilly (1991), O'Reilly et al. (2000), Parry (1965, 1981), Peters (1995), Rahman and Parry (1978) and Reynolds et al. (1994).

3. Concentration-dependent parameters

Commonly used room temperature magnetic parameters for assessing concentration are the mass specific values for susceptibility (χ), saturation magnetisation (σ_S), saturation remanent magnetisation (σ_{RS}) and anhysteretic remanent magnetisation (σ_{ARM}). The variation of χ, σ_{RS} and χ_{ARM} with grain size for magnetite, titanomagnetite, maghemite, hematite, goethite and pyrrhotite are shown in Fig. 2. The data is summarised in Table 1. In Fig. 2(a) magnetite, maghemite and titanomagnetite display the highest values of χ, with hematite and goethite the lowest values; pyrrhotite samples plot in between. The inset to Fig. 2(a) focuses on the variation of χ for magnetites up to 10 μm in size. Heider et al. (1996) propose that χ shows practically no grain-size dependence, with a mean value of 596 $\mu(\text{m}^3\,\text{kg}^{-1})$ for magnetite. O'Reilly (1984) based on general principles, and Dearing (1999) to schematise measured data, propose a peak in χ corresponding to the largest SP grains superimposed on a grain size independent base line. The magnetite samples shown on the inset to Fig. 2(a) follow the schematic diagram of Dearing (1999), but also include samples larger than SP with high χ values. The three samples with high χ values between 0.2 and 0.55 μm in size are all from Dunlop and Argyle (1997). The authors speculate that the high values would be due to particle elongation. Detailed grain-size determinations have not been published for greigite hence the absence of greigite data in the corresponding figures. From the small amount of published data, however, it follows that greigite on average has χ values between those observed for pyrrhotite and titanomagnetite (Table 1).

In Fig. 2(b) goethite samples display the lowest σ_{RS} values, with hematite slightly higher. Magnetite, titanomagnetite, maghemite and pyrrhotite all display similar high values of σ_{RS} (likewise greigite according to Table 1). Pyrrhotite has a saturation magnetisation, σ_S, five times lower than magnetite. Therefore, it is interesting to note that they both have similar σ_{RS} values (Fig. 2(b)). The coincidence is due to pyrrhotite having fewer domains than magnetite for a given grain size, thus behaving more SD-like with higher σ_{RS}/σ_S (or M_{RS}/M_S) ratios. σ_S is another parameter, which could be used as a concentration indicator, however relatively few authors actually determine and publish values of it (cf. Table 1). The values given for σ_S of magnetite in Table 1 are from Hartstra (1982b); they are slightly lower than the normally quoted value of 92 $\text{A m}^2\,\text{kg}^{-1}$ (e.g. Maher and Thompson, 1999) because the samples were not entirely saturated during measurement as noted by Vlag

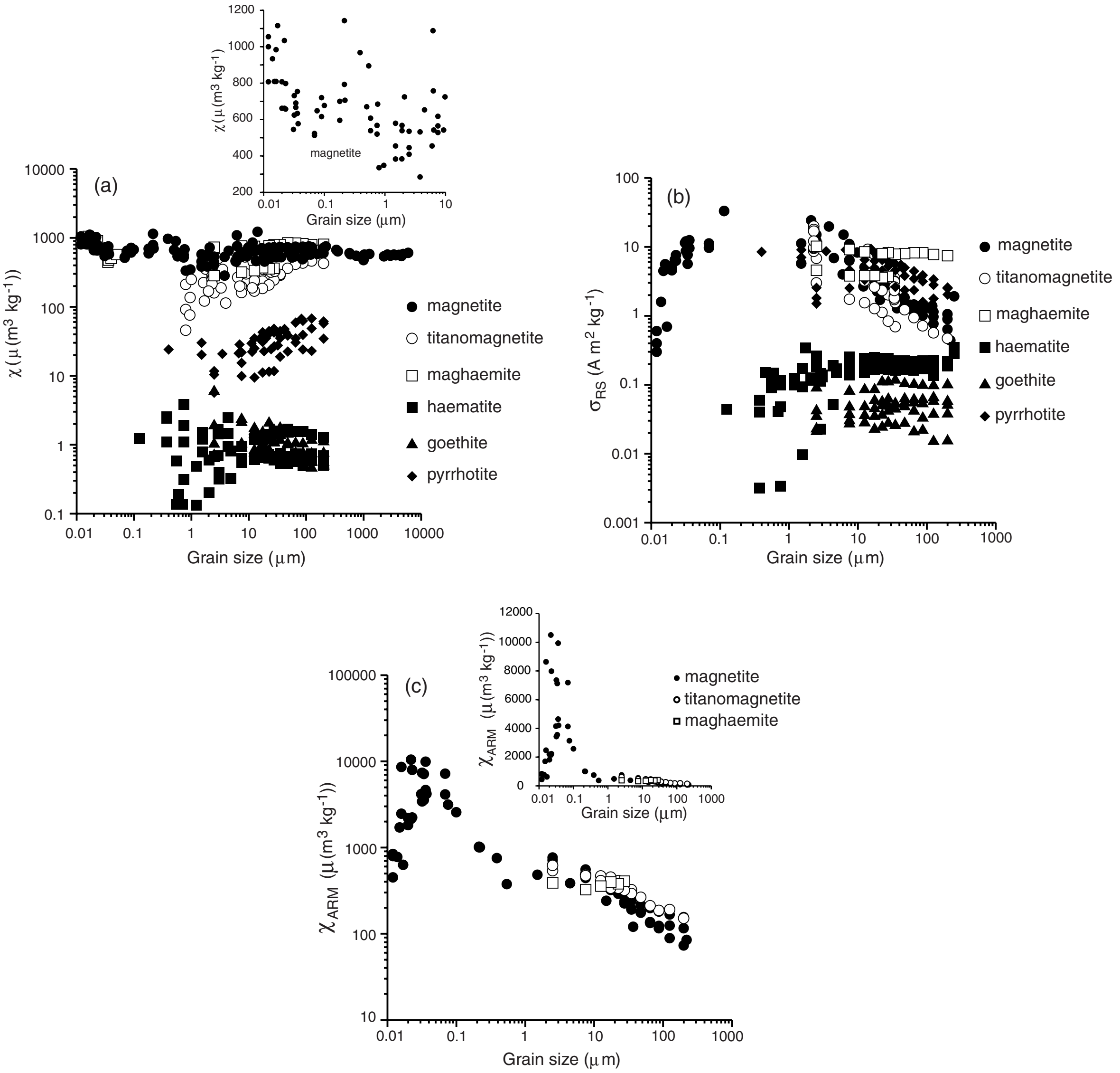

Fig. 2. Variation of concentration-dependent parameters (a) susceptibility, χ, (b) saturation remanent magnetisation, σ_{RS}, and (c) susceptibility of the anhysteretic remanent magnetisation, χ_{ARM}, with grain size for magnetite, titanomagnetite, maghemite, hematite, goethite and pyrrhotite. The data are from Bailey and Dunlop (1983), Clark (1984), Dankers (1978, 1981), Day et al. (1977), Dearing et al. (1996), de Boer and Dekkers (1996), Dekkers (1988, 1989), Dekkers and Linssen (1989), Dunlop (1971), Dunlop and Argyle (1997), Hartstra (1982a,b,c), Heider et al. (1996), Maher (1988), Menyeh and O'Reilly (1991), O'Donovan et al. (1986), O'Reilly et al. (2000), Parry (1965, 1981) and Rahman and Parry (1978). The inset to (a) focuses on the variation of χ for magnetites up to 10 μm. The inset to (c) shows the variation of χ_{ARM} on a linear scale.

et al. (1996) who dispersed again all fractions of two of Hartstra's samples and published different coercivity and remanence ratios.

In Fig. 2(c) the variation of the susceptibility of ARM (χ_{ARM}) is plotted for magnetite, titanomagnetite and maghemite (the only published ARM data found; Dankers, 1978; Dunlop and Argyle, 1997; Hartstra, 1982a,c; Maher, 1988; Parry, 1981). χ_{ARM} has been plotted to take into account the different bias fields used in ARM acquisition. For magnetite, ARM is proportional with the bias field for bias fields [$\mu_0 \times$field] up to approximately 0.1 mT, in higher bias fields ARM values start to level off slightly. The field at which the non-linearity becomes apparent is dependent on magnetic concentration; higher concentrations are linear over larger field ranges. Tauxe (1993) with reference to Sugiura (1979) showed that for a marine carbonate (i.e. low magnetic concentration) non-linearity was apparent even for values of [$\mu_0 \times$field] as low as 0.1 mT (the bias field recommended by King et al. (1982)). The peak alternating field is also of importance. Not all minerals (including some magnetites) will be saturated in a peak alternating field [$\times\mu_0$] of 100 mT. Future investigation of log-normal fitting of a full ARM acquisition curve could allow the maximum ARM to be calculated. In Fig. 2(c) similar χ_{ARM} values are observed for magnetite, titanomagnetite and maghemite for

Table 1
Summary of data in Figs. 1–4

Parameter or ratio	Mineral						
	Magnetite	Titano-magnetite	Maghemite	Hematite	Goethite	Pyrrhotite	Greigite
	Av. (no.) [min → max]						
χ (10^{-6} $m^3\,kg^{-1}$)	674 (98) [285 → 1233]	422 (86) [46 → 806]	632 (17) [283 → 845]	0.97 (63) [0.13 → 3.83]	1.17 (59) [0.46 → 5.92]	32.1 (54) [5.7 → 67.5]	108 (12) [26 → 194]
χ_{FD} (%)	2.4 (23) [0.4 → 11]	No data	No data	No data	No data	No data	No data
σ_{RS} ($A\,m^2\,kg^{-1}$)	5.3 (82) [0.3 → 33.1]	5.2 (43) [0.5 → 19.9]	6.8 (17) [3.6 → 10.2]	0.18 (95) [0.003 → 0.35]	0.052 (60) [0.015 → 0.12]	5.0 (54) [1.6 → 9.3]	5.4 (11) [0.8 → 12.3]
σ_S ($A\,m^2\,kg^{-1}$)	62.2 (12) [50.3 → 89.5]	16.0 (10) [11.2 → 21.9]	63.6 (11) [61.5 → 65.6]	0.28 (18) [0.093 → 0.47]	0.22 (58) [0.02 → 0.59]	13.3 (54) [3.5 → 21.0]	14.1 (12) [3.1 → 29.2]
χ_{ARM} (10^{-6} $m^3\,kg^{-1}$)	1673 (72) [73 → 10510]	362 (19) [151 → 623]	376 (6) [324 → 408]	No data	No data	No data	608 (12) [242 → 916]
σ_{RS}/χ (10^3 $A\,m^{-1}$)	11.3 (97) [0.3 → 80.8]	21.0 (53) [1.1 → 120]	11.0 (17) [9.2 → 16.1]	261 (63) [1.2 → 783]	57.4 (59) [5.9 → 212]	209 (64) [4.5 → 997]	70.7 (41) [11.2 → 174]
σ_{RS}/χ_{ARM} (10^3 $A\,m^{-1}$)	7.7 (67) [0.5 → 23.1]	5.02 (19) [2.15 → 11.0]	10.4 (6) [9.2 → 11.8]	No data	No data	No data	8.6 (11) [2.6 → 15.7]
χ_{ARM}/χ	2.37 (72) [0.11 → 13.7]	1.64 (19) [0.35 → 3.92]	1.12 (6) [1.02 → 1.37]	No data	No data	No data	5.6 (27) [1.8 → 9.8]
$(B_0)_C$ (10^{-3} T)	9.8 (101) [0.1 → 34.3]	20.8 (87) [2.0 → 158]	7.2 (11) [6.4 → 9.0]	268 (18) [4 → 520]	217 (60) [25 → 890]	36.9 (64) [9.8 → 97.2]	42.6 (67) [10.0 → 71.3]
$(B_0)_{CR}$ (10^{-3} T)	24.4 (185) [8.0 → 69.5]	41.4 (96) [8.5 → 213]	20.8 (17) [16.9 → 31.0]	318 (96) [30 → 821]	1972 (60) [500 → 4100]	45.3 (64) [10.0 → 124.5]	67.1 (78) [37.0 → 94.8]
$(B_0)_{CR'}$ (10^{-3} T)	30.8 (58) [10 → 63]	30.3 (25) [8.5 → 138]	30.7 (15) [25.0 → 46.5]	270 (82) [28 → 769]	3386 (60) [1200 → 6999]	52.6 (45) [16.4 → 134.0]	75.0 (12) [41.4 → 96.8]
$(B_0)_{CR}/(B_0)_C$	9.1 (96) [1.4 → 65]	3.2 (87) [1.2 → 6.9]	2.62 (11) [2.54 → 2.67]	1.76 (13) [1.28 → 2.19]	19.1 (60) [2.5 → 66.7]	1.27 (63) [0.55 → 2.29]	1.74 (62) [1.21 → 5.09]
$(B_0)_{CR}/(B_0)_{CR'}$	0.68 (58) [0.44 → 0.95]	0.77 (25) [0.54 → 1.0]	0.66 (15) [0.65 → 0.68]	1.05 (82) [0.79 → 1.24]	0.65 (60) [0.25 → 1.65]	0.80 (45) [0.54 → 1.22]	0.79 (12) [0.73 → 0.89]
M_{RS}/M_S	0.12 (89) [0.0005 → 0.4]	0.19 (63) [0.01 → 0.53]	0.13 (11) [0.12 → 0.16]	0.58 (18) [0.43 → 0.85]	0.32 (58) [0.06 → 1.0]	0.36 (64) [0.02 → 0.58]	0.45 (67) [0.18 → 0.69]

The average value (number of samples on which the average value is based) and the [minimum and maximum values] for magnetic parameters and ratios for seven magnetic minerals are given.

grains larger than 1 μm. For magnetite grains less than 1 μm in size a large peak in χ_{ARM} values is observed, which corresponds to the single domain (SD) state grain-size range where the anhysteretic remanent magnetisation is maximum (Fig. 2(c) inset). In the log–log plot (Fig. 2(c)) an exponential trend from ~50 nm upwards emerges, similar to the trend in coercivity values versus grain size. Greigite which is not shown in the figure, has average χ_{ARM} values greater than those for titanomagnetite and maghemite (cf. Table 1).

Of all the minerals considered, maghemite displays little variation in magnetic parameters with grain size (most apparent in Fig. 2(b)). de Boer and Dekkers (1996) attribute the lack of variation to the presence of both maghemite and hematite intimately intergrown in each fraction. In the large grains the maghemite parts are only up to ~20 μm in size, thus grains larger than 20 μm are agglomerates of maghemite/hematite intergrowths. Natural (and synthetic) maghemite will always be very small in grain size. Thermodynamics prescribe that the topotactic reaction by which maghemite forms from magnetite only supports the cation-deficient spinel structure in maghemite in small grains, smaller than approximately 1 μm. Larger grains invert to hematite, the thermodynamically stable Fe_2O_3 phase (de Boer, 1999 and references therein). Maghemites as large as observed by de Boer and Dekkers (1996) are rare; the maghemite structure is presumably stabilised by aluminium substitution.

Of the three parameters shown in Fig. 2, χ_{ARM}, displays the greatest grain-size dependence. Susceptibility (Fig. 2(a)) shows the least grain-size dependent trend and is therefore possibly the best parameter for assessing concentration. The trends displayed in the figures refer to those of the magnetic minerals, i.e. the matrix contribution to the susceptibility can be ignored or is corrected for. However, it is important to be aware of paramagnetic and diamagnetic contributions to χ. In many marine and other sediments, they can constitute up to 50% or more of the measured χ. High paramagnetic and diamagnetic components would also lead to low frequency dependent susceptibility, χ_{FD}, values. Therefore, low χ values should be considered in conjunction with the high-field slope of a hysteresis loop to assess the paramagnetic and diamagnetic contribution to the total susceptibility.

4. Parameters and ratios indicating domain state

By dividing two concentration-dependent parameters (e.g. those in Fig. 2) the resulting ratios are concentration-independent—at least at first order—and can be interpreted as indicating grain-size trends or mineralogical differences (e.g. in Fig. 1). Selected parameters, which are reputed to give an indication of domain state, are shown in Figs. 3 and 4. The grain-size variation of the ratios σ_{RS}/χ, σ_{RS}/χ_{ARM} and χ_{ARM}/χ is shown in Fig. 3. For grains larger than 1 μm (titano)magnetite display decreasing σ_{RS}/χ ratios with increasing grain size (Fig. 3(a)). However very small grains of magnetite appear to have similar σ_{RS}/χ ratios to those of very large magnetite grains. Pyrrhotite and hematite display the highest values of σ_{RS}/χ. Hematite displays no apparent trend with grain size. Pyrrhotite follows a trend similar to magnetite, but for any given grain size has σ_{RS}/χ ratios up to one hundred times higher than magnetite. Greigite has an average σ_{RS}/χ value greater than magnetite, but less than pyrrhotite (not plotted, see Table 1). In Fig. 3(b) and (c) values of σ_{RS}/χ_{ARM} less than 1 kA m^{-1} and values of χ_{ARM}/χ greater than 5 are indicative of the presence of very small magnetite grains. Greigite has a comparable average value of σ_{RS}/χ_{ARM} to (titano)magnetite and maghemite, but a higher average value of χ_{ARM}/χ (cf. Table 1).

The variation of the remanent coercivity with grain size is shown in Fig. 4(a). For grains larger than 1 μm, all minerals (except goethite) display a decrease in remanent coercivity with increasing grain size i.e. the grains are becoming increasingly MD. Goethite displays the opposite trend, where an increase in remanent coercivity is found with increasing grain size. This would indicate that goethite cannot be described as a classical antiferromagnetic material. Likely surface magnetic moments and their increased coupling in larger grains play an important role. For magnetite grains less than

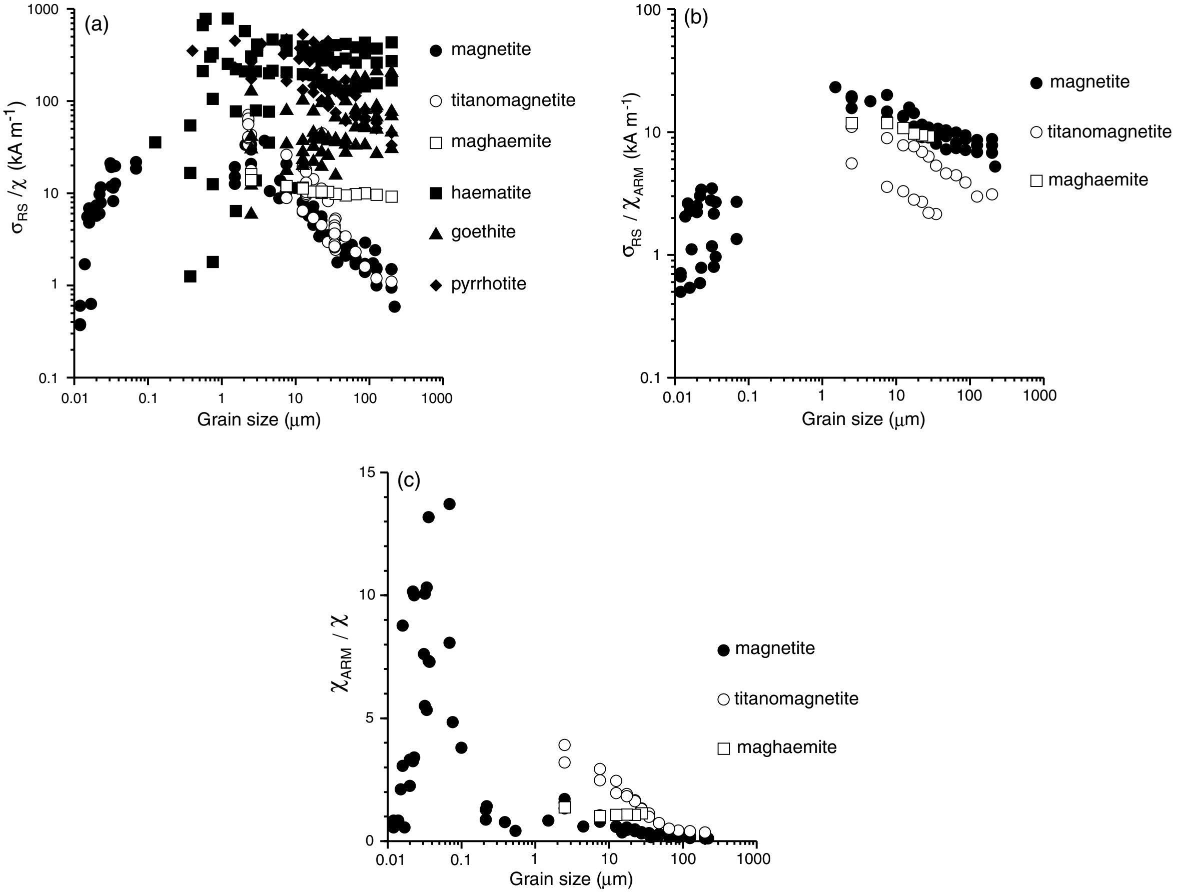

Fig. 3. Variation of the ratios (a) σ_{RS}/χ, (b) σ_{RS}/χ_{ARM}, and (c) χ_{ARM}/χ with grain size for magnetite, titanomagnetite, maghemite, hematite, goethite and pyrrhotite. The data are from Bailey and Dunlop (1983), Clark (1984), Dankers (1978, 1981), de Boer and Dekkers (1996), Dekkers (1988), Dekkers (1989), Dekkers and Linssen (1989), Dunlop (1971), Dunlop and Argyle (1997), Hartstra (1982a,b), Maher (1988), Menyeh and O'Reilly (1991), O'Reilly et al. (2000), Parry (1965, 1981) and Rahman and Parry (1978).

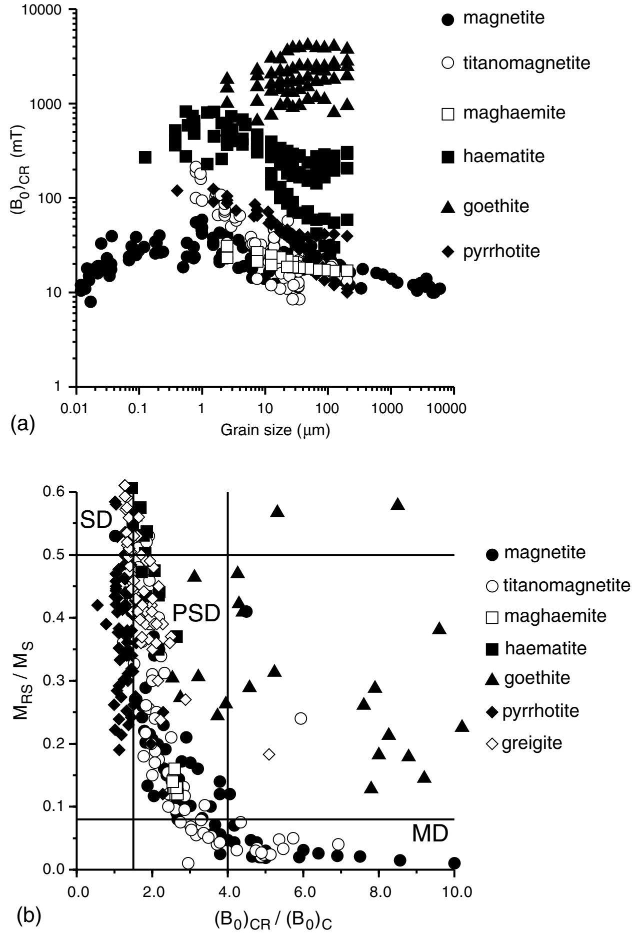

Fig. 4. (a) Variation of the remanent coercivity with grain size for magnetite, titanomagnetite, maghemite, hematite, goethite and pyrrhotite. (b) The Day et al. (1977) biplot of M_{RS}/M_S versus $(B_0)_{CR}/(B_0)_C$ for magnetite, titanomagnetite, maghemite, hematite, goethite, pyrrhotite and also greigite. The data are from Bailey (1975), Bailey and Dunlop (1983), Clark (1984), Dankers (1981), Day et al. (1977), de Boer and Dekkers (1996), Dekkers (1988), Dekkers (1989), Dekkers and Linssen (1989), Dekkers and Schoonen (1996), Dunlop (1971), Dunlop (1986), Hartstra (1982b,c), Heider et al. (1996), King et al. (1983), Levi and Merrill (1978), Maher (1988), Menyeh and O'Reilly (1991), Moskowitz et al. (1988), O'Donovan et al. (1986), O'Reilly et al. (2000), Özdemir and Banerjee (1982), Parry (1965, 1981), Peters (1995), Rahman and Parry (1978), Roberts (1995), Snowball (1995, 1997a,b) and Stanjek et al. (1994).

1 μm in size the remanent coercivity decreases with decreasing grain size, i.e. approaching 'true superparamagnetic (SP)' grains, which have zero remanence and zero coercivity by definition, for a given relaxation time that is smaller than the measurement time of the instrument used. Thus, low remanent coercivity values can suggest either a SP component or MD grains.

Fig. 4(b) shows the frequently used biplot of Day et al. (1977), which by combining hysteresis parameters suggests a graphical method for distinguishing the domain states SD to MD. Day et al. (1977) biplot was originally based on titano(magnetite) data. Theoretical aspects of the Day et al. (1977) biplot, including the influence of SP grains, are discussed by Dunlop (2002). Here the focus is on the applicability of the biplot to other mineralogies. The general trend for magnetite is followed by many of the other magnetic minerals, including hematite and greigite (Fig. 4(b)). Goethite, due to its hysteresis properties described above, displays no evident trend on the Day et al. (1977) biplot, instead the samples scatter to the right of the magnetite trend. Pyrrhotite, although following the general trend, appears to be shifted slightly to the left compared to the other minerals.

Figs. 3 and 4 show, overall, that domain state determination for well-characterised samples is not so straightforward. For many magnetic parameters small SP grains display properties similar to large MD grains. For the best interpretation in terms of domain state knowledge of the magnetic mineralogy is required. χ_{FD} values (cf. Table 1) can help to discriminate SP from MD grains. There seems to be a maximum observational limit on χ_{FD} of ~15% (Dearing et al., 1996). This limit has been argued to be the consequence of finite particle volume and coercivity distributions (Eyre, 1997; Worm, 1998). In cases of a very narrow grain-size distribution χ_{FD}'s of over 30% have been observed (Worm and Jackson, 1999). Muxworthy (2001) calculated decreases of up to 50% in χ_{FD} for weakly interacting samples compared to the non-interacting state of the same grain distributions. Magnetic interaction effectively reduces the blocking volume that is frequency dependent (Muxworthy, 2001). So, for environmental samples that may show variable grain size distributions and variable amounts of interaction, it is hard to tie the amount of frequency dependency in susceptibility to a specific grain size. If noticeable frequency dependence is observed the best label from a grain size view point would be a distribution that peaks in the 'large SP' size range.

5. Conclusions

1. Remanent acquisition coercivities and the ratio of the saturation remanent magnetisation to susceptibility are useful for assessing mineralogy. Very high remanent acquisition coercivities distinguish goethite from hematite. For the low coercivity minerals, pyrrhotite has high saturation remanent magnetisation to susceptibility ratios, whereas very low values of the ratio are characteristic of (titano)magnetite.
2. Of the three magnetic parameters susceptibility, saturation remanent magnetisation and susceptibility of anhysteretic remanent magnetisation, susceptibility showed the least dependence on grain size. Susceptibility is therefore, perhaps, the best parameter for assessing magnetic concentration in environmental samples, assuming uniform mineralogy and consideration of paramagnetic and diamagnetic components.

Table 2
Details of the conversions used to standardise all data into mass specific SI units

Magnetic parameter	Conversion	Notes	References
Susceptibility	Multiply by $4\pi/1000$	Conversion from χ_{CGS} to χ_{SI}	Dankers (1978), Dunlop (1971)
	Divide by density and multiply by $4\pi/1000$	Conversion from κ_{CGS} to χ_{SI}	Bailey and Dunlop (1983), Clark (1984), Day et al. (1977), Parry (1981), Rahman and Parry (1978)
	Multiply by 100, divide by density and multiply by $4\pi/1000$	Volume compensation and conversion from κ_{CGS} to χ_{SI}	Parry (1965)
	Divide by density	Conversion from κ_{SI} to χ_{SI}	Dunlop and Argyle (1997), Heider et al. (1996), O'Donovan et al. (1986)
	Divide by 1000	Assumed given values are in SI units per gram rather than kilogram	Hartstra (1982a,b,c)
Magnetisation	Multiply by 1	Conversion from σ_{CGS} to σ_{SI}	Clark (1984), Dunlop (1971)
	Divide by density	Conversion from $\mathbf{M}_{CGS}$ or $\mathbf{M}_{SI}$ to σ_{SI}	Bailey and Dunlop (1983), Parry (1981), Rahman and Parry (1978)
	Multiply by 100 and divide by density	Volume compensation and conversion from $\mathbf{M}_{CGS}$ to σ_{SI}	Parry (1965)
Coercivities	Multiply by $4\pi \times 10^{-7}$	Conversion from $\mathbf{H}_{SI}$ to $\mathbf{B}_{SI}$	Dekkers (1988), Dekkers and Linssen (1989), Menyeh and O'Reilly (1991), O'Donovan et al. (1986)
	Divide by 10^4	Conversion from $\mathbf{H}_{CGS}$ to $\mathbf{B}_{SI}$	Bailey and Dunlop (1983), Clark (1984), Day et al. (1977), Dunlop (1986), Özdemir and Banerjee (1982), Parry (1965, 1981), Rahman and Parry (1978)

3. Investigation of domain state is complicated by very small and large grains displaying similar magnetic properties. On the Day et al. (1977) biplot a general trend is followed by the minerals (titano)magnetite, maghemite, hematite and greigite. Pyrrhotite is shifted slightly to the left of the general trend. Goethite samples scatter to the right of the general trend.
4. More data is required to complement the existing data set to formulate observed experimental trends in terms of 'expected' grain size trends especially for very fine grains in the SP/SD grain-size realm of mineralogies other than magnetite.

Acknowledgements

The work was conducted under the programme of the Vening Meinesz Research School of Geodynamics (VMSG), Utrecht University (The Netherlands). Thoughtful reviews of Adrian Muxworthy and an anonymous reviewer were appreciated. Funding was provided by the European Union through the MAG-NET network (contract number FMRX980247) in the frame of the CORDIS RTD-PROJECTS.

Appendix A. Standardisation of data

In some of the referenced publications the data were given in 'CGS' units, whilst in other publications volume specific parameters were quoted. Table 2 lists the conversion factors used in converting the data into mass specific SI units.

References

Bailey, M.E., 1975. The magnetic properties of pseudo-single domain grains. M.Sc. thesis, University of Toronto.

Bailey, M.E., Dunlop, D.J., 1983. Alternating field characteristics of pseudo-single-domain (2–14 μm) and multi-domain magnetite. Earth Planet. Sci. Lett. 63, 335–352.

Clark, D.A., 1984. Hysteresis properties of sized dispersed monoclinic pyrrhotite grains. Geophys. Res. Lett. 11 (3), 173–176.

Dankers, P.H.M., 1978. Magnetic properties of dispersed natural iron-oxides of known grain-size. Ph.D. thesis, University of Utrecht.

Dankers, P.H.M., 1981. Relationship between median destructive field and remanent coercive forces for dispersed magnetite, titanomagnetite and haematite. Geophys. J. R. Astr. Soc. 64, 447–461.

Day, R., Fuller, M., Schmidt, V.A., 1977. Hysteresis properties of titanomagnetites: grain-size and compositional dependence. Phys. Earth Planet. Int. 13, 260–267.

Dearing, J.A., 1999. Magnetic susceptibility, In: Walden, J., Oldfield, F., Smith, J. (Eds.), Environmental Magnetism; a practical guide. Quaternary Research Association Technical Guide No. 6, London.

Dearing, J.A., Dann, R.J.L., Hay, K., Lees, J.A., Loveland, P.J., Maher, B.A., O'Grady, K., 1996. Frequency-dependent susceptibility measurements of environmental materials. Geophys. J. Int. 124, 228–240.

de Boer, C.B., 1999. Rock-magnetic studies on hematite, maghemite and combustion-metamorphic rocks. PhD thesis, University of Utrecht.

de Boer, C.B., Dekkers, M.J., 1996. Grain-size dependence of the rock-magnetic properties for a natural maghemite. Geophys. Res. Lett. 23, 2815–2818.

Dekkers, M.J., 1988. Magnetic properties of natural pyrrhotite part 1: behaviour of initial susceptibility and saturation-magnetization-related rock-magnetic parameters in a grain-size dependent framework. Phys. Earth Planet. Int. 52, 376–393.

Dekkers, M.J., 1989. Magnetic properties of natural goethite—I. Grain-size dependence of some low- and high-field related rock-magnetic parameters measured at room temperature. Geophys. J. 97, 323–340.

Dekkers, M.J., Linssen, J.H., 1989. Rockmagnetic properties of fine-grained natural low-temperature haematite with reference to remanence acquisition mechanisms in red beds. Geophys. J. Int. 99, 1–18.

Dekkers, M.J., Schoonen, M.A.A., 1996. Magnetic properties of hydrothermally synthesized greigite (Fe_3S_4)—I. Rock magnetic parameters at room temperature. Geophys. J. Int. 126, 360–368.

Dunlop, D.J., 1971. Magnetic properties of fine-particle hematite. Ann. Geophys. 27 (3), 269–293.

Dunlop, D.J., 1986. Hysteresis properties of magnetite and their dependence on particle size: a test of pseudo-single-domain remanence models. J. Geophys. Res. 91 (B9), 9569–9584.

Dunlop, D.J., 2002. Theory and application of the Day plot (M_{rs}/M_s versus H_{cr}/H_c) 1. Theoretical curves and tests using titanomagnetite data, J. Geophys. Res. 107(B3), 10.1029/2001JB000486.

Dunlop, D.J., Argyle, K.S., 1997. Thermoremanence, anhysteretic remanence and susceptibility of submicron magnetites: Nonlinear field dependence and variation with grain size. J. Geophys. Res. 102, 20199–20210.

Dunlop, D.J., Özdemir, Ö., 1997. Rock Magnetism—Fundamentals and Frontiers. Cambridge Studies in Magnetism. Cambridge University Press, Cambridge, UK.

Eyre, J.K., 1997. Frequency dependence of magnetic susceptibility for populations of single-domain grains. Geophys. J. Int. 129, 209–211.

Hartstra, R.L., 1982a. A comparative study of the ARM and Isr of some natural magnetites of MD and PSD grain size. Geophys. J. R. Astr. Soc. 71, 497–518.

Hartstra, R.L., 1982b. Grain-size dependence of initial susceptibility and saturation magnetization-related parameters of four natural magnetites in the PSD-MD range. Geophys. J. R. Astr. Soc. 71, 477–495.

Hartstra, R.L., 1982c. Some rockmagnetic parameters for natural iron-titanium oxides. Ph.D. thesis, University of Utrecht.

Heider, F., Zitzelsberger, A., Fabian, K., 1996. Magnetic susceptibility and remanent coercive force in grown magnetite crystals from 0.1 μm to 6 mm. Phys. Earth Planet. Int. 93, 239–256.

Heslop, D., Dekkers, M.J., Kruiver, P.P., Van Oorschot, I.H.M., 2002. Analysis of isothermal remanent magnetisation acquisition curves using the expectation–maximisation algorithm. Geophys. J. Int. 148, 58–64.

Hunt, C.P., Moskowitz, B.M., Banerjee, S.K., 1995. Magnetic properties of rocks and minerals. In: Ahrens, T.J. (Ed.), Rock Physics and Phase Relations. A Handbook of Physical Constants. American Geophysical Union, pp. 189–204.

King, J., Banerjee, S.K., Marvin, J., Ozdemir, O., 1982. A comparison of different magnetic methods for determining the relative grain size of magnetite in natural materials: some results from lake sediments. Earth Planet. Sci. Lett. 59, 404–419.

King, J.W., Banerjee, S.K., Marvin, J., 1983. A new rock-magnetic approach to selecting sediments for geomagnetic paleointensity studies: application to paleointensity for the last 4000 years. J. Geophys. Res. 88 (B7), 5911–5921.

Levi, S., Merrill, R.T., 1978. Properties of single-domain, pseudo-single-domain, and multidomain magnetite. J. Geophys. Res. 83, 309–323.

Maher, B.A., 1988. Magnetic properties of some synthetic sub-micron magnetites. Geophys. J.R. Astr. Soc. 94, 83–96.

Maher, B.A., Thompson, R. (Eds.), 1999. Quaternary Climates, Environments and Magnetism. Cambridge University Press, Cambridge, UK.

Menyeh, A., O'Reilly, W., 1991. The magnetization process in monoclinic pyrrhotite (Fe_7S_8) particles containing few domains. Geophys. J. Int. 104, 387–399.

Moskowitz, B.M., Frankel, R.B., Flanders, P.J., Blakemore, R.P., Schwartz, B.B., 1988. Magnetic properties of magnetotactic bacteria. J. Magn. Magn. Mater. 73, 273–288.

Muxworthy, A.R., 2001. Effect of grain interactions on the frequency dependence of magnetic susceptibility. Geophys. J. Int. 144, 441–447.

O'Donovan, J.B., Facey, D., O'Reilly, W., 1986. The magnetization process in titanomagnetite ($Fe_{2.4}Ti_{0.6}O_4$) in the 1–30 μm particle size range. Geophys. J.R. Astr. Soc. 87, 897–916.

O'Reilly, W., 1984. Rock and Mineral Magnetism. Blackie, Glasgow, UK.

O'Reilly, W., Hoffmann, V., Chouker, A.C., Soffel, H.C., Menyeh, A., 2000. Magnetic properties of synthetic analogues of pyrrhotite ore in the grain size range 1–24 μm. Geophys. J. Int. 142, 669–683.

Özdemir, Ö., Banerjee, S.K., 1982. A preliminary magnetic study of soil samples from west-central Minnesota. Earth Planet. Sci. Lett. 59, 393–403.

Parry, L.G., 1965. Magnetic properties of dispersed magnetite powders. Philos. Mag. 11 (110), 303–312.

Parry, L.G., 1981. The influence of fine structures on the remanence of multidomain particles of magnetite and titanomagnetite. Phys. Earth Planet. Int. 26, 63–71.

Peters, C., 1995. Unravelling magnetic mixtures in sediments, soils and rocks. Ph.D. thesis, University of Edinburgh.

Peters, C., Thompson, R., 1998. Magnetic identification of selected natural iron oxides and sulphides. J. Magn. Magn. Mater. 183, 365–374.

Rahman, A.A., Parry, L.G., 1978. Titanomagnetites prepared at different oxidation conditions: hysteresis properties. Phys. Earth Planet. Int. 16, 232–239.

Reynolds, R.L., Tuttle, M.L., Rice, C.A., Fishman, N.S., Karachewski, J.A., Sherman, D.M., 1994. Magnetization and geochemistry of greigite-bearing Cretaceous strata, Noth Slope Basin, Alaska. Am. J. Sci. 294, 485–528.

Roberts, A.P., 1995. Magnetic properties of sedimentary greigite (Fe_3S_4). Earth Planet. Sci. Lett. 134, 227–236.

Snowball, I.F., 1995. Mineral magnetic and geochemical properties of Holocene sediments and soils in the Abisko region of northern Sweden. Ph.D. thesis, Lund University.

Snowball, I.F., 1997a. The detection of single-domain greigite (Fe_3S_4) using rotational remanent magnetization (RRM) and the effective gyro field (Bg): mineral magnetic and palaeomagnetic applications. Geophys. J. Int. 130, 704–716.

Snowball, I.F., 1997b. Gyroremanent magnetization and the magnetic properties of greigite-bearing clays in southern Sweden. Geophys. J. Int. 129, 624–636.

Stanjek, H., Fassbinder, J.W.E., Vali, H., Wagele, H., Graf, W., 1994. Evidence of biogenic greigite (ferrimagnetic Fe_3S_4) in soil. Eur. J. Soil Sci. 45, 97–103.

Sugiura, N., 1979. ARM, TRM and magnetic interactions: concentration dependence. Earth Planet. Sci. Lett. 42, 451–455.

Tauxe, L., 1993. Sedimentary records of relative paleointensity of the geomagnetic field: theory and practice. Rev. Geophys. 31, 319–354.

Thompson, R., Oldfield, F., 1986. Environmental Magnetism. George Allen and Unwin, London.

Vlag, P.A., Rochette, P., Dekkers, M.J., 1996. Some additional hysteresis parameters for a natural (titano)magnetite with known grain size. Geophys. Res. Lett. 23, 2803–2806.

Walden, J., Oldfield, F., Smith, J., 1999. Environmental magnetism—a practical guide. Quaternary Research Association Technical guide No. 6, London, UK.

Worm, H.-U., 1998. On the superparamagnetic-stable single domain transition for magnetite, and frequency dependence of susceptibility. Geophys. J. Int. 133, 201–206.

Worm, H.-U., Jackson, M.J., 1999. The superparamagnetism of Yucca Mountain tuff. J. Geophys. Res. 104, 25415–25425.