How environmental magnetism can enhance the interpretational value of grain-size analysis: A time-slice study on sediment export to the NW African margin in Heinrich Stadial 1 and Mid Holocene

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A R T I C L E  I N F O

Article history:
Received 22 November 2013
Received in revised form 25 March 2014
Accepted 8 April 2014
Available online 21 April 2014

Keywords:
Environmental magnetism
Grain-size analysis
Aeolian sediment transport
Senegal River
Provenance
Northwest Africa

A B S T R A C T

Sediment dynamics in limnic, fluvial and marine environments can be assessed by granulometric and rock-magnetic methodologies. While classical grain-size analysis by sieving or settling mainly bears information on composition and transport, the magnetic mineral assemblages reflect to a larger extent the petrology and weathering conditions in the sediment source areas. Here, we combine both methods to investigate Late Quaternary marine sediments from five cores along a transect across the continental slope off Senegal. This region near the modern summer Intertropical Convergence Zone is particularly sensitive to climate change and receives sediments from several aeolian, fluvial and marine sources. From each of the investigated five GeoB sediment cores (494–2956 m water depth) two time slices were processed which represent contrasting climatic conditions: the arid Heinrich Stadial 1 (−15 kyr BP) and the humid Mid Holocene (−6 kyr BP). Each sediment sample was split into 16 grain-size fractions ranging from 1.6 to 500 μm. Concentration and grain-size indicative magnetic parameters (susceptibility, SIRM, HIRM, ARM and ARM/IRM) were determined at room temperature for each of these fractions. The joint consideration of whole sediment and magnetic mineral grain-size distributions allows to address several important issues: (i) distinction of two aeolian sediment fractions, one carried by the north-easterly trade winds (40–63 μm) and the other by the overlying easterly Harmattan wind (10–20 μm) as well as a fluvial fraction assigned to the Senegal River (−10 μm); (ii) identification of three terrigenous sediment source areas: southern Sahara and Sahel dust (low fine-grained magnetite amounts and a comparatively high haematite content), dust from Senegalese coastal dunes (intermediate fine-grained magnetite and haematite contents) and soils from the upper reaches of the Senegal River (high fine-grained magnetite content); (iii) detection of partial diagenetic dissolution of fine magnetite particles as a function of organic input and shore distance; (iv) analysis of magnetic properties of marine carbonates dominating the grain-size fractions 63–500 μm.

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1. Introduction

The Northwest African realm is particularly sensitive to climate change as it is located off the Sahel, a transition zone between the humid tropics and the arid subtropics (White, 1983). Changes in precipitation as well as in direction and intensity of low- and high-altitude winds have large impact on the origin and properties of sediments deposited off Northwest Africa (Fig. 1). The region is also intriguing as it is under the influence of two aeolian transport systems, the north-easterly trade winds and the overlying easterly Harmattan, as well as one fluvial transport system, the Senegal River. As the predominant climatic factors and sediment transport mechanisms are essentially understood thanks to earlier palaeoclimatic studies (e.g. Koopmann, 1981; Sarnthein et al., 1981; Mulitza et al., 2008; Itambi et al., 2009; Just et al., 2012a), this study area is very appropriate to test whether grain-size specific environmental analysis has the potential to deliver further insights.

Sediment grain-size distribution analyses provide important information to reconstruct past and present sediment transport dynamics and depositional conditions. Grain-size methods have been successfully applied in a vast number of sedimentological studies covering all climate zones and settings (Allen, 1965; Folk, 1966; Holland and Elmore, 2008; Poizot et al., 2008). In contrast, sediment provenance as well as pedogenic and post-depositional alteration state are primarily derived from sediment petrological observations. Environmental magnetism is in principle an efficient lithostratigraphic method to explore these questions; it is based on the marker potential of sedimentary magnetic minerals and the variation of magnetic properties through time and space (e.g. Liu et al., 2012). The approach is particularly useful to reconstruct weathering
Fig. 1. Set of maps showing the study area under the modern climate of Northwest Africa and the adjacent Atlantic with main terrigenous sediment sources and transport systems (aeolian/fluvial): (a) Mean Aerosol Index (non-dimensional) for the years 2006–2013 (freely available OMI/AURA data from http://mirador.gsfc.nasa.gov/), values \( < 2.0 \) mark areas with frequent sand and dust storm events due to convection of air masses; (b) Accumulated precipitation for boreal winter (December-January-February; DJF) of the year 2011/12 (freely available TRMM data from http://mirador.gsfc.nasa.gov/), the approximate location of the Intertropical Convergence Zone (ITCZ) is shown in a simplified way as dashed black line with the tropical rain belt being located to its south over land, north-easterly and south-easterly trade winds are shown as grey arrows (compiled and drawn after Ruddiman, 2001), the main aeolian dust sources in map (a) are displayed as grey areas; (c) Accumulated precipitation for boreal summer (June-July-August; JJA) of the year 2012, the tropospheric circulation is displayed schematically to visualise the mechanism of dust uplift from main sources of NW Africa (after Nicholson, 2009), the ITCZ (area between dashed black lines) and the tropical rain belt are quasi-independent features, areas of air convection are marked by blue circles with a central dot, areas of air subsidence are marked by grey crossed circles, the African Easterly Jet (AEJ) above the convection cell of the ITCZ is shown as broad orange arrow, the Tropical Easterly Jet (TEJ) is symbolised by orange and black circles, the convergence (CONV) and the convection of air at the ITCZ and the tropical rain belt are shown as blue arrows, the flow of the trade winds and air subsidence after divergence (DIV) are marked by grey arrows. (d) Bathymetry (GEBCO 0.5 min grid) and locations of 5 gravity cores obtained during the METEOR research-cruise M65/1 (Mulitza et al., 2006) and sampled within this project; (e) Vertical cross-section of the tropospheric circulation cells averaged for 10° W to 10° E during boreal summer (drawn after Nicholson, 2009), the latitudinal scale is identical to (c), AEJ and the Tropical Easterly Jet (TEJ) are symbolised by orange and black circles, the convergence (CONV) and the convection of air at the ITCZ and the tropical rain belt are shown as blue arrows, the flow of the trade winds and air subsidence after divergence (DIV) are marked by grey arrows.
conditions in sediment source areas (e.g. Maher, 1986; Lyons et al., 2010) and to detect syn- or post-depositional Fe-mineral diagenetic alteration (e.g. Karlin and Levi, 1985; Hesse, 1994; Robinson et al., 2000; Fu et al., 2008). Magnetic Fe-oxide and sulphide minerals are relatively low in species number and concentration. However, even trace concentrations (ppm level) can be quantified by rock-magnetic methods and provide clues as to their magmatic, metamorphic, pedogenic, biogenic or diagenetic origin (e.g. Franke et al., 2007).

Like other proxies of sediment composition, grain-size and rock-magnetic records reflect the combined impact of multiple sediment sources and transport modes. Provenance and transport vary as function of changing climatic and weathering conditions. These convolved contributions are the rule in many "palaeoestudies"; individual contributions can only be made visible by separating transport, source and climate signatures – a nearly hopeless task if only bulk sediment properties are known. Rock-magnetic properties are dependent on the mineralogy of the magnetic particles and on their concentration, but also on their grain size. Therefore, grain-size fractionation prior to rock-magnetic analysis enables constraining magnetic mineralogy and concentration effects to certain grain-size fractions while enhancing their diagnostic potential.

So far, environmagnetic methods have been rarely utilised in the analysis of sieved sediment grain-size fractions. Previous studies mainly dealt with riverine and coastal environments (e.g. Maher et al., 2009). The magnetic characterisation of soils from different tributaries plays an important role in demonstrating changing land-use patterns or to detect variations in precipitation and surface erosion (Oldfield et al., 1985; Hatfield and Maher, 2009). A coastal magnetic study by Oldfield and Yu (1994) was able to separate magnetic provenance signals in the coarse silt and sand fractions from a bacterial magnetite signal associated to the clay fraction. Magnetic provenance “fingerprinting” has been used to describe along- and across-shelf sediment dynamics (e.g. Hatfield et al., 2010). In their extensive magnetogranulometric study on modern North Atlantic sediments, Hatfield et al. (2013) could distinguish Greenlandic and Icelandic terrestrial provenances. Both deliver magnetite particles of similar sizes but with different abundances within the clastic grain-size spectrum. These results illustrate the well-known rock-magnetic phenomenon that clastic and magnetogranulometric grain sizes often differ substantially as many magnetite particles are inclusions in larger silicate host grains (Hounslow and Maher, 1996; Otofuji et al., 2000; Hounslow and Morton, 2004).

Our time-slice study intends to lead somewhat further: By testing a grain-size specific rock-magnetic approach in (i) a glacial arid and (ii) a post-glacial humid climate, we aim to resolve the impacts of individual sediment sources as well as of various transport systems. In particular, we examine whether or not magnetic particles are being enriched or depleted in specific grain-size ranges by these afore-mentioned factors. It is also essential to test whether or not the grain sizes of magnetic particles co-vary with clastic grain size and, if so, how strong this link is for various magnetic mineral associations and provenances. We will also try to derive palaeoclimatic information from the preservation of fine magnetic particles and pinpoint magnetite dissolution throughout the clastic grain-size range.

Therefore, we have resampled five well-studied Late Quaternary sediment gravity cores from a depth transect across the continental slope off Senegal (Fig. 1a,d). For each core, we identified the stratigraphic position of the arid Heinrich Stadal 1 (H1S1, ~15 kyr BP) and of the Mid Holocene (MH, ~6 kyr BP) at the end of the African Humid Period (Mulitza et al., 2008). The arid H1S1 at Northwest Africa was caused by the slowdown of the Atlantic Meridional Overturning Circulation (e.g. Mulitza et al., 2008 and references therein). This slowdown, due to inputs of large amounts of melt water from the high-latitude continental ice shields, led to cooling of the North Atlantic and increased the temperature gradient between the higher and lower latitudes. This in turn shifted the tropical rain belt southwards and amplified the northern easterly trade winds. In contrast, the MH (and the entire African Humid Period) was characterised by weakened trade winds and by a more northern position of the tropical rain belt compared to today. A gradual increase in boreal summer insolation led to high precipitation over the Sahara which was almost entirely vegetated in those times (e.g. deMenocal et al., 2000).

The sediment samples of both time slices from the five cores were separated into 16, on average ~0.5 µm wide, grain-size fractions ranging from 1.6 to 500 µm. This yielded 160 grain-size fractions of which a set of rock-magnetic parameters was determined at room temperature. Subsequently, we interpreted grain-size dependent trends in the magnetic susceptibility as well as in isothermal, anhysteretic and high-field magnetic remanence intensities and compared them to values obtained for the bulk samples.

2. Sedimentary and environmental setting

2.1. Tropospheric circulation of Northwest Africa

The scene of this study, the northern part of the Sahel, is a savannah landscape and semi-desert experiencing yearly precipitations from 0 to 500 mm (Li et al., 2004). While annual and seasonal precipitation controls weathering at terrestrial sediment sources, wind intensities and seasonally predominant wind directions determine the main transport systems. All of them experience systematic climate-driven changes as shown by the marine archives off-coast Senegal (Michel, 1973 and references therein; Diaster-Haass, 1976; Mulitza et al., 2008; Iambi et al., 2009).

Some traditional, but rather simplified explanations for seasonal precipitation and wind patterns of Northwest Africa (e.g. Dettwiller, 1965; Germain, 1968), however, are still prevailing. As the tropospheric circulation is of palaeoclimatic relevance for the interpretation of the H1S1 and MH data, we give a short introduction to modern views based on the seminal work of Nicholson (2009). Further, a review paper by Zhang et al. (2006) on this topic is interesting reading.

During boreal winter (December-January-February; DJF), Northwest Africa and the northern Senegalese continental margin are under influence of boreal north-easterly trade winds (Fig. 1b). These winds are directed from the North-Saharan high-pressure cell towards the tropical low-pressure trough (Intertropical Convergence Zone; ITZC) located at around 5° N over the Atlantic and central Africa (e.g. Sultan and Janicot, 2003 and references therein). At the ITZC, the north-easterly trade winds from the Sahara and Sahel converge with austral south-easterly trade winds transporting oceanic moisture. Their joint updraft leads to cooling at higher tropospheric levels and to abundant daily rainfalls (Parker et al., 2005).

During boreal summer (June-July-August; JJA) the area of maximum insolation shifts northwards (Fig. 1c). At low-tropospheric levels, the austral south-easterly trade winds arise at the equator-ward side of the South Atlantic high-pressure cell and cross over the equator. They are deflected northeastwards by Coriolis forces and by the enhanced surface low-pressure cell (ITZC) over Northwest Africa (Xie and Saiki, 1999). Although a significant part of the moisture for precipitation comes from the Atlantic Ocean (African coast between 5° and 10° N; Fig. 1c), the main portion of the moisture for convection and on-land precipitation is derived from local evaporation (Nicholson, 2009). The responsible force for the convection of air masses within the “tropical rain belt” over land (the zone of maximum summer precipitation, mainly at 5°–15° N; Fig. 1c,e) is the cross-equatorial pressure gradient. The intensity of this deep convection is primarily influenced by the dynamics of the African Easterly Jet (AEJ) and the Tropical Easterly Jet (TEJ) rather than by wind or moisture convection (Nicholson, 2009). Simultaneously with the northward shift of the tropical rain belt, the ITZC shifts northward to 15°–23° N at the surface, reaching the latitudes of the northern Sahel and southern Sahara (Tetzlaff and Peters, 1986; Sultan and Janicot, 2003; Nicholson, 2009). At the ITZC, the south-westerly
monsoon winds merge with the north-easterly trade winds and the overlying easterly Harmattan (Fig. 1;c.e; latter not marked here for clarity) (Sultan and Janicot, 2003; Nicholson, 2009), also referred to as "Saharan Air Layer" in literature (Tetzlaff and Peters, 1986; Middleton and Goudie, 2001: Stuut et al., 2005 and references therein). This rather shallow air convection cell is responsible for precipitation over the northern Sahel and southernmost Sahara particularly during wet years, being quasi-independent from the tropical rain belt (Nicholson, 2009).

2.2. Aeolian transport systems and related terrestrial sediment sources

As the ITZC creates little precipitation during mean boreal summers, the concerned area provides the largest source for aeolian dust worldwide (Fig. 1a,c) (e.g. Mahowald et al., 2006; Roberts et al., 2011). During boreal winter conditions, the north-easterly trade winds as well as the overlying easterly Harmattan are the major suppliers of aeolian dust towards the Senegalese continental margin. The north-easterly trade winds pick up relatively coarse particles from dunes of the western Sahara (Table 1) and transport them at low-tropospheric levels of up to 1000–1500 m (Fig. 1a,b) (Koopmann, 1981; Sarnthein et al., 1981; Grousset et al., 1998; Middleton and Goudie, 2001). These particle distributions have modal values between 40 and 70 μm (Grousset et al., 1998); only ca. 5–13% of the grains are covered with haematite (Koopmann, 1981 and references therein).

The north-easterly trade winds are overlain by the easterly Harmattan, which is strongest during summer months due to an enhanced AEJ (Fig. 1c.e). The Harmattan and AEJ transport the largest part of aeolian sediments from Sahel and southern Sahara to the Atlantic (e.g. Middleton and Goudie, 2001; Caquineau et al., 2002; Holz et al., 2004; Stuut et al., 2005). Both climatic features are located at mid-tropospheric levels (1500–5000 m) with highest aeolian dust concentrations at ~3000 m (Tetzlaff and Peters, 1986). The dust is mainly picked up from southern Sahara and Sahel Zone dry lands (Fig. 1a) by vertical atmospheric turbulence, i.e. dust outbreaks (Fig. 1e) (e.g. Holz et al., 2004; Stuut et al., 2005). Main aeolian sediment sources are the Bôddé Depression in Chad (Fig. 1a), the Southern Sahara of Algeria, northern Mali and western Morocco (Middleton and Goudie, 2001; Caquineau et al., 2002 and references therein). Sahel dust sources between 10’ and 16’ N are second most important sources (Tetzlaff and Peters, 1986).

Although these southern Sahara and northern Sahel sediment sources show loessoidal characteristics with grain-size modes between 16 and 60 μm, the transported aeolian dust is typically between 10 and 30 μm (Table 1) (Koopmann, 1981 and references therein; Holz et al., 2004; Stuut et al., 2005; Lyons et al., 2012). Particles ~30 μm fall out after few tens of kilometres, creating residual Sahara and Sahel sediment covers (Lyons et al., 2010, 2012). Aeolian dust particles <16 μm can be created by crumbling of coarser particles during salting. Between 35% and 75% of the Harmattan and AEJ sediments are haematite-coated (Table 1) (Koopmann, 1981 and references therein). Besides abundant haematite coatings, these sediments also contain high amounts of pedogenically formed maghaemite. This makes them more strongly magnetic than the northern and north-western Sahara sediments. These important insights were established with the help of grain-size based rock and environmental magnetic studies of Lyons et al. (2010, 2012), which were performed on dust samples collected along north-south trending transects across the whole Sahara and Sahel.

In the coastal region, the lower zone of the Harmattan shows increased turbulence with the underlying moist and cool north-easterly trade winds. This leads to a marked and relatively abrupt gravitational fall-out of its coarsest dust grains with sizes >40 μm (Koopmann, 1981) originating from coastal dunes. The grain size of these aeolian sediments decreases with shore distance. Particles <12 μm can reach as far as the Caribbean (Prospero et al., 1981; Middleton, 1985; Swap et al., 1996), but fine aeolian particles can also be found in significant concentrations at the continental margin off Senegal (Koopmann, 1981; Sarnthein et al., 1981; Grousset et al., 1998; Middleton and Goudie, 2001). These particles act as nuclei for moisture condensation and are easily washed out by rain. Clay particles often aggregate already in the air column. Aggregation is even more intense after the saline sea surface is reached. The settling process within the ocean can further be enhanced by suspension feeders digesting the aggregates and bonding them to larger faecal particles. Such aggregates obtain settling characteristics of fine sandy quartz and can reach the ocean floor within a few days to weeks (Koopmann, 1981 and references therein; Grousset et al., 1998 and references therein), where they disintegrate. This sedimentation process is therefore relatively fast and the grain-size distributions along our sediment core transect (Fig. 1a,d) should mainly mirror these aeolian transport dynamics and not be biased by ocean currents at the continental slope (Koopmann, 1981; Sarnthein et al., 1981; Mulitza et al., 2008; Itambé et al., 2009).

2.3. Fluviatile system and related terrestrial sediment sources

A major contributor of silty to clayey sediments to the northern Senegalese continental margin is the Senegal River (Fig. 1a–d; Table 1) (Michel, 1973; Lesack et al., 1984; Nizou et al., 2010 and references therein). Its average yearly runoff is 641 m³ s⁻¹, with a yearly sediment load between 1.2 and 2.8 Mt yr⁻¹ and a mean suspended solid concentration of 196–252 mg l⁻¹ (Gac and Kane, 1986). The Senegal drainage basin contains the oldest rocks at its upper reaches, the Fouta Djallon and Mading Plateaus, which constitute the strongly magnetic Precambrian basement rocks of the West African Shield. They are overlain by Palaeozoic schists and quartizes, sedimentary sandstones together with few schists and dolomites, as well as basalts and rhyolites intersected by diorite and granite intrusions.

The middle and lower reaches of the Senegal drainage basin consist mainly of Mesozoic-Tertiary sandstone formations. Only the Senegal River delta is covered by Quaternary alluvial deposits. From an environmentally view, the igneous rocks of the upper drainage reaches are the most important source for weathering products like ferruginous and ferralitic soils that contain high concentrations of ferrimagnetic and antiferromagnetic Fe-oxides. The upper drainage reaches of the Senegal River in Mali also appear to be a prominent source for fine-grained, pedogenically formed, magnetite and maghaemite (Lyons et al., 2010). Their concentrations in the soils increase southwards with

<table>
<thead>
<tr>
<th>Sediment source</th>
<th>Transport system</th>
<th>Mode size carried to NW Africa</th>
<th>Magnetite/maghaemite</th>
<th>Haematite</th>
</tr>
</thead>
<tbody>
<tr>
<td>Western Sahara</td>
<td>NE trade winds</td>
<td>40–70 μm</td>
<td>Main appearance as inclusions in host grains +/–</td>
<td>Coatings on quartz grains ++</td>
</tr>
<tr>
<td>Southern Sahara/Sahel</td>
<td>Harmattan</td>
<td>10–30 μm</td>
<td>Pedogenic individual grains +/–</td>
<td>Coatings on quartz grains ++</td>
</tr>
<tr>
<td>Fouta Djallon/Manding Plateaus</td>
<td>Senegal River</td>
<td>~6 μm</td>
<td>Pedogenic/igneous individual grains ++</td>
<td>Coatings on quartz and clays +</td>
</tr>
</tbody>
</table>

+/– rare + common ++ very common.

Table 1

Properties of terrigenous source sediments relevant for our grain-size based rock-magnetic study off Senegal (from earlier studies mentioned in Sections 2.2 and 2.3: Koopmann, 1981; Sarnthein et al., 1981; Grousset et al., 1998; Middleton and Goudie, 2001; Lyons et al., 2010, 2012).
higher precipitation (Lyons et al., 2010). When the river reaches the saline ocean water, the suspended particles settle and the dissolved Fe-complexes precipitate. The availability of such syn-sedimentary precipitates results in the formation of olive-green to grey Fe-rich and strongly paramagnetic marine clay deposits (Odin, 1988; Itambi et al., 2009).

2.4. Hydrography and sedimentology at the continental margin off northern Senegal

The continental margin off northern Senegal and Mauritania between Cape Verde and Cape Blanc (15°–21° N) is characterised by a relatively narrow continental shelf of 40–60 km width (Fig. 1a,d) and a 100–200 m deep shelf break (McMaster and Lachance, 1969; Wynn et al., 2000). The continental slope has a width of ~80 km and descends at an angle of ~1.5° down to 2500 m water depth.

The oceanoigraphy at the continental margin between Cape Verde and Cape Blanc has been described by Hagen (2001) and Stramma et al. (2005). During boreal summer, the inner and middle parts of the continental shelf (0–100 m) are under influence of the Tropical Surface Water from a northward-flowing coastal current driven by south-westerly monsoon winds. Simultaneously, the northward-flowing South Atlantic Central Water overflows the outer shelf and the upper slope (100–500 m). During boreal winter, this subsurface water is upwelling onto the entire shelf and driven southwards by the north-easterly trade winds, displacing the Tropical Surface Water southwards. This seasonally clockwise or anti-clockwise circulating shelf waters have a significant influence on the sediment transport to the slope (e.g., Nizou et al., 2010 and references therein). During boreal summer, coastal currents transport sediments to the north up to Cape Blanc and then down-slope and partially backward to the south (Koopmann, 1981 and references therein). Thus, the shelf and upper continental slope between the Senegal River mouth and Cape Blanc are generally free of silty and clayey sediments, which are then found in higher concentrations at the continental rise (Barusseau et al., 1988; Nizou et al., 2010). This part of the continental shelf is mainly covered with medium or fine (aeolian) sands (~75% of sediment composition) with less than 10% of silt and 10% of clay (McMaster and Lachance, 1969 and references therein).

Between the Senegal River mouth and Cape Verde to its south (Fig. 1d), a high proportion of very fine fluvial material (Barusseau et al., 1988; Nizou et al., 2010) is deposited besides coarser quartz particles during boreal winter. As these sediments do not reach areas south of Cape Verde, they must be diverted off-shelf and down-slope toward our sediment core transect. Here, the inner shelf deposits (0–50 m water depth) show the largest grain-size range and most heterogeneous distributions of the shelf deposits (Lézine and Chateauanneuf, 1991; Nizou et al., 2010). Locally, they are described as sands, muddy sands or sandy muds. The mid shelf (50–100 m) is mainly covered with sandy muds and the outer shelf (100–200 m) with muddy sands. The surface sediments of the adjacent continental slope (200–2500 m) consist of muddy silts (silt comprises 40% of terrigenous fraction with minor amounts of sandy and clayey grains) and the sediments of the continental rise (2500–4500 m) are composed of clayey muds to muddy clays (Koopmann, 1981).

3. Materials and methods

3.1. Materials

This study is based on five sediment gravity cores collected at the continental margin off northern Senegal during METEOR cruise M65/1 (Mulitza et al., 2006). These are GeoB 9513-3, 9512-5, 9510-1, 9508-5 and 9506-1 (from shallow to deep, Fig. 1a,d; Table 2) and come from water depths between 494 m and 2956 m following an ESE-WNW trending transect across the slope. Each core was sampled at two levels (20–30 ml each) corresponding to the time slices HS1 (~15 kyr BP) and the MH (~6 kyr BP).

The age model of the deepest core GeoB 9506-1 was based on a benthic δ18O curve and correlated to the already dated cores GeoB 9516-4 (Itambi et al., 2009) and MD 95-2042 (Shackleton et al., 2004). Core GeoB 9508-5 has its independent 14C age model (Mulitza et al., 2008). The HS1 and MH time slices of the remaining three sediment cores were defined by correlation of rock-magnetic, element and colour data to the GeoB 9506-1 age model (Itambi et al., 2009). In all cores, the arid HS1 level is represented by reddish-brown sandy muds of distinctly aeolian origin (Just et al., 2012a). The colour is due to haematite coatings on quartz particles and clay minerals. During the African Humid Period of the MH, climate conditions enabled the deposition of dark-green silty muds, containing an important contribution from fine-grained Senegal River sediments (e.g., Odin, 1988).

3.2. Methods

3.2.1. Wet sieving

Before sieving, the samples were freeze-dried to determine their dry bulk masses of 18–52 g while avoiding alteration and aggregation. About 20% of the dry sediment was removed for bulk sediment measurements. The other 80% of the dry bulk sample were suspended in 100 ml of demineralised water. 3.5 g of tetra-sodiumdiphosphate was added to prevent creation of aggregates in the clay fraction (following e.g., Barbanti and Bothner, 1993). The sediment was then further diluted with 200 ml demineralised water and ultrasonically dispersed in a BRANSON Sonifier 450 homogeniser for 3–5 minutes.

Wet sieving was performed with the sifting machine Fritsch analysette 3 with ISO-3310 sieves. The maximum water discharge of the glands was ~1.5 l per minute at ~2 bar water pressure. Sieving was done with vibration amplitude of 2.0 mm. The sieving time was 10 minutes in total with shake intervals of 3 s, being followed by a pause of 1 s. Ten grain-size fractions from >500 μm to >20 μm (Table 3) were obtained using 10 stacked sieves.

3.2.2. Settling method

A settling method served to partition the <20 μm effluent into 6 further fractions (Table 3). The settling times for each targeted fraction (i.e. maximum diameter) were calculated from Stokes’ law as shown by Oldfield and Yu (1994) and Lamb (2006) assuming a sediment density equal to that of quartz. Two sequential runs per settled grain-size fraction were performed (Table 3) to reduce contamination by finer particles which settle simultaneously with the larger target-size particles from the starting suspension.

The settling was started with retention of the coarsest particles to obtain as much mass as possible from each fraction. Hence, settling of particles finer than the target grain size is a disadvantage of this separation method and leads to an underestimation of the finer grain-size fractions obtained later. Therefore, a more representative (main) grain size was calculated, which was qualitatively defined by the largest mass proportion within each settled fraction applying again Stokes’

Table 2 Locations and water depths of five sampled GeoB gravity cores recovered from the continental slope off Senegal, Northwest Africa.

<table>
<thead>
<tr>
<th>Core (GeoB)</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Water depth</th>
<th>Shore distance</th>
</tr>
</thead>
<tbody>
<tr>
<td>9506-1</td>
<td>15° 36.60’ N</td>
<td>18° 21.01’ W</td>
<td>2956 m</td>
<td>163 km</td>
</tr>
<tr>
<td>9508-5</td>
<td>15° 29.90’ N</td>
<td>17° 56.88’ W</td>
<td>2378 m</td>
<td>118 km</td>
</tr>
<tr>
<td>9510-1</td>
<td>15° 20.09’ N</td>
<td>17° 39.20’ W</td>
<td>1566 m</td>
<td>85 km</td>
</tr>
<tr>
<td>9512-5</td>
<td>15° 20.31’ N</td>
<td>17° 22.02’ W</td>
<td>795 m</td>
<td>53 km</td>
</tr>
<tr>
<td>9513-3</td>
<td>15° 19.11’ N</td>
<td>17° 17.64’ W</td>
<td>494 m</td>
<td>45 km</td>
</tr>
</tbody>
</table>
law to mathematically simulate the two settling runs. Target (maximum) and corresponding main grain sizes are listed in Table 3. The calculated main grain sizes were validated by incident light microscopy on a set of selected grain-size fractions. In the following, we show grain-size distributions with respect to the main grain size of quartz as most dominant mineral in terms of mass. The rock-magnetic parameters are also shown against quartz equivalent grain size to facilitate comparison of the entire range of measured parameters within this study. As the individual sieve fractions vary extremely in volume, it was more practical to normalise susceptibility and remanences by sample mass.

### Table 3
Grain-size fractions obtained by wet sieving and settling listed by their maximum and main grain size based on the density of quartz. Shown are settling times for first and second runs. The main grain size was assumed by calculating the most abundant grain size with respect to finer grains being previously caught in coarser fractions.

<table>
<thead>
<tr>
<th>Maximum-grain size [µm]</th>
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### Table 4
Masses and relative abundances of grain-size fractions from the Heinrich Stadial 1 time slice.

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3.2.3. Sample preparation for magnetic measurements

To enable meaningful low-field magnetic susceptibility (MS) and remanence measurements, all fractions with their widely varying mass (Tables 4 and 5) were set either in molten paraffin that was moulded into spherules or fixed in gelatine capsules. This prevents grain rotation in the applied magnetic fields and centres the small amounts of material in the standard 6.2 cm³ plastic sample cubes. Both procedures yielded measurement results of comparable quality. We favoured the gelatine capsule method, since it is faster and also more convenient for further analyses (e.g. microscopy). Hard-gelatine capsules of size Ø 23.3 mm; Ø 8.2 mm) of high purity were used. To avoid grain rotation, the sediment in one half of the capsule was firmly compressed with the outside of the reversibly inserted second half. Both halves were fixed with Scotch Tape and slid into a drinking straw of an appropriate diameter. To centre the sediment in the standard 6.2 cm³ plastic cubes, the drinking straw was trimmed to the necessary length. The mass of each individual gelatine capsule was measured to subtract its diamagnetic contribution from the measured MS.

3.2.4. Magnetic susceptibility and remanence measurements

MS measurements were performed at 920 Hz and 320 A m⁻¹ rms-field strength using a KAPPABRIDGE KLY-2 instrument with a sensitivity of 0.04 × 10⁻⁶ SI. All measurements were corrected for sample holder, casing and stabiliser effects.

Measurements of the anhysteretic and isothermal remanent magnetisations (ARM and IRM respectively) were performed with a fully automated 2G ENTERPRISES755R DC-SQUID pass-through magnetometer at the Marine Geophysics Section, University of Bremen (Germany). The sensitivity of this equipment is 1.0 × 10⁻¹² Am² corresponding to 0.1613 µA m⁻¹ for a sample of 6.2 cm³.

Table 5
Masses and relative abundances of grain-size fractions from the Mid Holocene time slice.

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Table 6
Set of commonly applied rock and environmental magnetic parameters (e.g. Liu et al., 2012) used in this study.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Definition</th>
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</thead>
<tbody>
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<td>MS</td>
<td>Magnetic susceptibility as mass-specific (χ) or volume-specific (κ): ratio between induced magnetisation of a material and the applied magnetic field; measure for (relative) concentration of ferrimagnetic, antiferromagnetic, paramagnetic and diamagnetic minerals.</td>
</tr>
<tr>
<td>IRM</td>
<td>Isothermal Remanent Magnetisation: magnetisation applied stepwise in direct magnetic fields at constant room temperature.</td>
</tr>
<tr>
<td>SIRM</td>
<td>Saturation IRM: IRM acquired (in this study) at a 700 mT magnetic field, above which no remanence acquisition takes place due to maximum possible magnetisation; measure for (relative) joint concentration of ferrimagnetic magnetite and antiferromagnetic haematite.</td>
</tr>
<tr>
<td>HIRM</td>
<td>High coercive IRM: difference between SIRM and remanence acquired at a 300 mT field; measure for (relative) concentration of magnetically high-coercive haematite.</td>
</tr>
<tr>
<td>S-ratio</td>
<td>Saturation ratio: remanence obtained in a 300 mT direct field related to SIRM; measure for contribution of pure magnetite to entire sediment magnetisation; under reductive diagenetic alteration the value decreases.</td>
</tr>
<tr>
<td>ARM</td>
<td>Anhysteretic Remanent Magnetisation: applied in 40 µT direct field; based on relatively low intensity of the applied field and its fast parallel/anti-parallel alternation, ARM is dominantly carried by Single Domain magnetite particles (0.02–0.1 µm).</td>
</tr>
<tr>
<td>ARM/IRM</td>
<td>Ratio between remanences acquired in alternating and direct fields of 100 mT; measure for (relative) concentration of fine magnetite particles in relation to coarse; ratio increases with fining particles; with coarsening particles and under reductive diagenetic alteration of magnetite the ratio decreases.</td>
</tr>
</tbody>
</table>
During ARM measurement, the fractions were magnetised in a 0.1 T alternating magnetic field (biasing direct field of 40 μT) followed by demagnetisation in 15 steps. IRMs were measured after application of incrementally stronger direct current pulse fields up to 0.7 T in 21 steps. After reaching the internal pulse-coil maximum field of 0.7 T, the samples were treated as magnetically saturated and the corresponding IRM is labelled SIRM. Although the saturation of haematite is not complete (pure haematite requires saturation fields of 1–5 T), it was sufficient for the relative quantification purposes of the present study.

The environmagnetic methodology provides a range of sedimentary parameters to: (i) mark source areas; (ii) identify pathways of aeolian, fluvial and marine sediments and (iii) detect authigenic Fe-minerals formed by biomineralisation, abiotic precipitation and diagenetic processes. In this study, we mainly focus on the properties of magnetite and haematite, as the first has the strongest impact on the magnetic properties of the sediment and the second is the most abundant Fe-mineral by mass in our research area. The rock-magnetic parameters selected for our study and their interpretation is listed in Table 6. For detailed explanations, several review papers are available; the most recent review is Liu et al. (2012).

4. Results

Grain-size specific mass distribution curves and magnetic parameters for each sediment core and time slice are shown in joint displays in Fig. 2a–h. The HS1 curves are always blue hatched and the MH curves green solid. Bulk property values (where shown) are visualised by red triangles (HS1) and circles (MH) for comparison. Due to inconsistent settling and quasi-permanent abeyance of light clay minerals (due to electrostatic interactions), we decided not to analyse fractions <1.6 μm. As we do not expect any sediment-source signal being exclusively related by grain size, the relative quantification was performed by dividing the magnetic mineral fraction by its respective mass fraction, i.e. domain state.

4.1. Grain-size distributions of total sediment and of magnetic minerals

The grain-size distribution curves (Fig. 2a) of both time slices illustrate that most particles have grain sizes of 10–70 μm. HS1 sediments show distributions peaking between 40 and 70 μm, strongly skewed towards finer grain sizes. MH grain-size distributions tend to be more bimodal over the coarse silt (40–70 μm) and fine silt (10–20 μm) range. In HS1, the 40–70 μm mode remains dominant over the entire transect, while in MH the 10–20 μm mode becomes more prominent with shore distance and is dominant at the distal locations. Contrasting with the varying amounts of the 40–70 μm mode, the 10–20 μm mode shows no obvious changes between the two time slices. The MH grain-size distributions are also characterised by a higher content of fine particles (<10 μm), in particular at the deeper locations. Cumulative recoveries for the HS1 range from 75% at the deepest to 91% at the shallower location (Table 4). For the MH slices, the cumulative recoveries (40–84%) were lower due to the much higher clay contents (14–67% instead of 10–31%) (Table 5).

First assumptions about corresponding magnetic mineral grain-size distributions (Fig. 2b) were obtained by multiplying the mass-specific susceptibility χ of every grain-size fraction with its respective mass portion (Tables 4 and 5) and hereafter named as “mass-weighted”. Within the investigated grain-size range, the bulk MS resides mainly in fractions of 10–40 μm during HS1 and 10–20 μm during MH; finer fractions down to 1.6 μm also have noticeable influence. Fractions ≥63 μm do not appear to have significant impact on the bulk MS in both time slices. Similar findings are also observed for mass-weighted ARM$_{0.1}$ and SIRM$_{0.7}$ grain-size distribution curves, which are not shown here. All these distributions primarily reflect the behaviour of the mineral magnetite due to its prominent MS.

4.2. Grain-size specific magnetic mineral concentrations

If MS χ is displayed without relation to fraction mass (Fig. 2c), it represents the magnetic mineral proportion in each grain-size fraction for the given location and time slice. For grain sizes >40 μm, the MS values appear to be quite irregularly distributed. From 40 μm down to 10 μm grain size, MS increases systematically. For even finer grain sizes, the MS values tend to level off and decrease slightly. MS is systematically lower in all MH fractions, averaging ~20 × 10$^{-8}$ m$^3$ kg$^{-1}$, while HS1 values double or even triple. In HS1 fractions, MS slightly decreases with shore distance, while no such trend is observed for MH.

Likewise for both time slices, mass-specific SIRM values (Fig. 2d) are very low above 40 μm, except for the coarsest HS1 fractions (355–500 μm). SIRM takes much higher values below 40 μm and reaches its maximum at ~10 μm. At the three distal locations, the differences between MH and HS1 time slices are small, but a decline with shore distance can be observed. The two most proximal cores GeoB 9512-5 and 9513-3 show similar SIRM distributions during HS1 as in the other cores. However, their MH SIRM values are anomalously low in comparison to these outside cores. Fig. 2c,d shows similar trends which indicate that MS behaviour is governed by magnetic minerals. SIRM is not affected by paramagnetism and diamagnetism, but reflects magnetic particle size, i.e. domain state.

HIRM values (Fig. 2e) are low above 40 μm and increase sharply below that grain size, reaching maxima in the 10–20 μm fractions. They are about two times higher in HS1 as in MH. HIRM values >40 μm are low and irregularly distributed as in the case of MS. No trend with shore distance can be observed for the whole grain-size range.

ARM values (Fig. 2f) in grain-size fractions >40 μm are very low. Interestingly ARM values increase only in the <20 μm fractions. For HS1, ARM behaviour does not change much as function of shore distance. In contrast, MH trends differ vastly: At the two most proximal locations, fine magnetic particles seem to be nearly absent, while in the central core GeoB 95010-1 MH values are times those of HS1. The HS1 bulk values of MS, SIRM, HIRM and ARM are in most cases about half of the fractions’ maximum values. For MH, however, bulk values resemble more the high values of the finest fractions collected. As the masses of MH fractions <1.6 μm are larger than those of HS1 (cf. Tables 4 and 5), the formers have consequently a stronger magnetic influence on the bulk values. We are aware that a portion of fine magnetic particles was lost in the grain-size extraction procedure. Although fractions <1.6 μm may have a considerable influence on the bulk signal, in the present study they can be neglected as it is unlikely that they contain information about a separate sediment source not being already represented by the properties of coarser fractions.

4.3. Grain-size specific magnetic granulometry and mineralogy

Through all MH fractions, the two most proximal cores show rather constant and low ARM/IRM values, pointing towards diagenetic depletion of fine magnetite grains. Interestingly for the other cores, the magnetogranulometric proxy parameter ARM/IRM (Fig. 2g) does not

Fig. 2. The measured parameters for each grain-size fraction of the core transect are displayed for the Heinrich Stadial 1 (HS1) or Mid Holocene (MH) by blue triangle-marked lines with horizontal hatching or by green dotted lines with solid background, respectively. Bulk sample values are shown as red triangles (HS1) or red dots (MH); (a) Grain-size mass distributions are shown as the cumulative sum derivative (spline of second order) of sieved and settled fractions; (b) Mass-weighted susceptibility (representing mineral grain-size distribution) of both time slices was obtained by multiplying susceptibility χ with the mass portion of the respective grain-size fraction (relative to the cumulative recovered mass ≥1.6 μm); (c) MS χ, (d) SIRM$_{0.7}$, (e) ARM$_{0.1}$ and (f) HIRM$_{0.7}$ are all shown on a mass-specific basis; (g) Magnetogranulometric ARM$_{0.1}$/IRM$_{0.7}$ ratio compares relative particle size of magnetite with clastic particle size; (h) SIRM$_{0.7}$ displayed as indicator of magnetite to haematite content. A few values of coarse fractions lie outside the data range and are not displayed for better clarity.
show the anticipated linear grain-size dependence, but prominent bimodal characteristics. With one exception, all distribution curves have minima at 20 μm marking the coarsest magnetite crystal sizes. Towards finer and coarser clastic grain-sizes, ARM/IRM increases systematically in both time slices. The largest increase is observed in the MH sample of the central core GeoB 9510-1. No trend with shore distance is visible in HS1. The MH fractions of the three outer cores show a smaller ARM/IRM increase with shore distance.

S-ratio values (Fig. 2h) are relatively low, i.e. more haematite-dominated, between 10 and 63 μm and reach nearly 1.00, i.e. magnetite-dominated, for the coarser (>63 μm) fractions. The relative magnetite proportion also increases towards finer grain sizes below 10 μm. At the three most distal core locations, the MH fractions show less haematite predominance than the HS1 fractions. At the two most proximal core locations, MH grain-size fraction values indicate selective magnetite depletion with respect to haematite. For both time slices, there is no obvious trend with shore distance.

5. Interpretation and discussion

Before we can interpret measured trends in terms of provenance and palaeoclimate, we need to consider potential biasing factors of magnetic properties such as diagenetic dissolution of magnetite and dilution by biogenic carbonates. Afterwards, we aim to distinguish the two relevant aeolian transport systems (north-easterly trade winds and Harmattan) and deduce palaeoclimatic changes from sediment flux variations. Finally, we will draw attention to the sediment contributions by the Senegal River.

5.1. Diagenetic dissolution of fine-grained magnetite

Grain-size fractionation clearly facilitates to demonstrate the known effects of reduction and dissolution of sedimentary Fe-oxides during organic matter degradation and remineralisation (Karlin and Levi, 1983; Canfield and Berner, 1987; Canfield et al., 1992; Tarduno, 1994; Smirnov and Tarduno, 2000; Fu et al., 2008; Rowan et al., 2009). Magnetite reduction occurs primarily during the MH at the two proximal sites as documented by unusually low SIRM, ARM and ARM/IRM values (Fig. 2d,f,g). The two inner core sites are closer to the Senegal River mouth and receive a higher input of fresh terrestrial organic matter (Zarriess and Mackensen, 2010). Magnetite reduction primarily affects isolated grains due to their much larger specific surface area.

The impact of grain-size selective diagenesis on the magnetic characteristics is even better shown in crossplots in Fig. 3: While (a) highlights the intensity of diagenetic magnetite depletion within the clastic grain-size fractions, (b) points out the different magnetic “palaeoclimate trends” of MH and HS1, where the magnetite-depleted MH grain-size
fractions resemble more the characteristics of the HS1 fractions than those of unaltered MH fractions. Finally, Fig. 3c points out that selective diagenesis biases magnetogranulometric information.

5.2. Dilution by marine carbonates

Besides the gradual concentration increase of fine magnetite particles in fining clastic grain-size fractions < 20 μm (Fig. 4a), fractions ≥ 63 μm show enrichment of fine magnetite particles with no consistent trend. These unexpected magnetogranulometric ARM/IRM trends ≥ 63 μm have three potential reasons: (i) autochthonous benthic and planktonic foraminifera fragments occur as individual particles within this grain-size range. As these hollow carbonate shells are fragile, it is conceivable that fine-grained magnetites may have entered their interior during burial; (ii) fine-grained magnetic particles often occur as inclusions in large clastic particles of igneous origin (Hounslow and Maher, 1996; Otofuji et al., 2000; Hounslow and Morton, 2004); (iii) as magnetites are heavy minerals, they are rarely found in deep marine sediments as individual terrigenous particles of > 63 μm. Due to their higher density, fine heavy minerals are often incorporated into shells of agglutinated benthic foraminifera, which makes them less prone to current drifts (e.g. Makled and Langer, 2010).

As we observe diagenetic dissolution of magnetite throughout the entire MH grain-size range (Figs. 3 and 4a) at the two shallowest locations, only hypothesis (i) reasonably explains the inconsistent relative enrichment of fine magnetites versus coarse magnetites in fractions ≥ 63 μm. While fragile foraminifera can contain fine magnetites, they do not protect them from dissolution. Explanations (ii) and (iii) are still possible, but are likely of minor importance. Terrigenous quartz of igneous origin with particles of 63–500 μm is rarely transported to the continental slope and rise by either aeolian or fluvial systems. Agglutinating benthic foraminifera would be more appropriate as they explain at least the minor magnetic peaks at 63–90 μm (Fig. 2c–f). However, fine-grained magnetite inclusions in quartz grains or in agglutinating benthic foraminifera would be protected from selective diagenetic dissolution, while even coarser individual magnetites could be dissolved. Under such protection, fine inclusions would lead to an increase of the granulometric proxy ARM/IRM, which contradicts our observation throughout all altered grain-size fractions (Fig. 4a). Therefore, in the next sections, these relatively small fractions ≥ 63 μm will be considered as less relevant as well as less significant for the reconstruction of sediment sources and palaeoclimatic changes.

5.3. Aeolian sediment fluxes and sources

To carry out magnetic fingerprinting of distinct Northwest African dust species, we begin with identifying grain-size ranges with characteristic haematite-magnetite proportions. Dust particles transported by north-easterly trade winds should be found in the 40–63 μm mode (Table 1), where intermediate proportions of haematite-coated particles with respect to fractions > 63 μm and < 40 μm are observed (Fig. 4b). Particles transported by the Harmattan would be expected in the 10–20 μm mode, which contains the highest proportion of haematite and lowest of fine-grained magnetite in respect to other fractions (Fig. 4a). Although haematites occur as coatings on particles particularly in the Harmattan mode (Fig. 4b), it is likely that haematite particles in the < 10 μm fractions were also mainly transported by the Harmattan. The notion that the < 10 μm fractions are an extension of the Harmattan 10–20 μm mode is most clearly supported by the steady ARM/IRM trend between 1.6 and 20 μm in the HS1 time slice (Fig. 4a). The consistent fining of clastic and magnetic grain sizes indicates that fine magnetite grains seem to exist as individual particles unlike the haematite coatings. The ARM/IRM values of the trade-wind mode (40–63 μm) lie outside this trend. The ARM/IRM vs. SIRM cross-plots (Fig. 5a,b) also show that the magnetic characteristics of the coarse end of the Harmattan fraction (20–μm) are very similar to those of the < 20 μm dust. The 10 μm Harmattan fraction is indistinguishable from the properties of the < 10 μm fractions, while the trade-wind fraction of 40 μm clearly differs magnetically.

The Harmattan sediment load originates from the Sahel Zone (Fig. 1a), as it resembles the properties of the strongly magnetic Sahelian surface sediments (Figs. 4a,b and 5a,b; Table 1) (Koopmann, 1981; Lyons et al., 2010, 2012). Their haematite coatings are created by pedogenic processes during successive arid and humid seasons (Kämpf and Schwertmann, 1983; Schwertmann and Taylor, 1989). Consequently, the quartz grains of the arid southern and western Sahara show lower concentrations of haematite-coatings. Their magnetic properties (Figs. 4a,b and 5a,b) resemble those being transported by the north-easterly trade winds. More distal sediment sources for the trade-wind mode (40–63 μm) than the white coastal Senegalese dunes can be excluded, as particles with grain sizes of 40–63 μm can only be transported over 20–100 km (Koopmann, 1981 and references therein; Grousset et al., 1998).

Based on the grain-size distributions (Fig. 2a) as well as on the magnetomineralogy (Fig. 4b) and magnetogranulometry (Figs. 4a and 5a,b), we are now able to reconstruct palaeoclimatic impacts on the transport dynamics: During the arid HS1 in respect to the humid MH,
the north-easterly trade winds were strengthened, as seen in the higher portion of the 40–63 μm mode (Fig. 2b). While the sediments carried by the trade winds (40–63 μm) do not show any strong change neither in ARM/IRM (Fig. 4a) nor in HIRM (Fig. 4b) from HS1 to MH, there is a clearly visible shift in the Harmattan mode (10–20 μm). As, however, the 10–20 μm grain-size mode (Fig. 2a) does not shift or increase in proportion, we deduce that the Harmattan at these latitudes was not stronger during HS1 than during MH. Stronger HS1 trade winds and the rather constant wind velocities of the Harmattan are in agreement with earlier studies (e.g. Koopmann, 1981; Gasse and van Campo, 1994; Gasse, 2000; Mulitza et al., 2008; Tjallingii et al., 2008). During the MH, the Sahel Zone and the southern Sahara had a dense vegetation cover (denser than today’s) inhibiting haematite-coated particles from being uplifted by winds. This would raise the relative proportion of haematite-poor sediments from the southern Sahara within the Harmattan dust flux.

5.4. Fluvial sediment flux and sources

During the arid HS1, large parts of the shelf were subaerially exposed and covered by coast-line parallel dune fields acting as barriers for the Senegal River (Nicholson and Flohn, 1980; Koopmann, 1981; Nizou et al., 2010). Senegal River sediments with typical grain sizes of <10 μm (e.g. Koopmann, 1981) were therefore not exported to the Northwest African continental slope (Nicholson and Flohn, 1980; Koopmann, 1981; Gasse and van Campo, 1994; Gasse, 2000; Mulitza et al., 2008; Nizou et al., 2010). Terrigenous particles <10 μm (Fig. 2a) deposited at the northern Senegalese continental slope during this time, can consequently not be directly linked to the fluvial transport system. Although the continental slope underwent strong upwelling by intensified north-easterly trade winds during HS1 (e.g. Zarriss and Mackensen, 2010; Just et al., 2012a), we can exclude resuspension of former fluvial silts and clays at the upper continental slope and shelf. Remobilised silts would show a down-slope directed concentration increase (defined as sortable silt of 10–63 μm; e.g. McCave et al., 1995), as they are easiest to be remobilised by currents after deposition. This scenario can be ruled
out as no distortion of a consistent shore-distance (down-slope) trend is
detectable in the grain-size distributions (Fig. 2a).

In contrast, due to the increased precipitation during the African Humid Period, the MH sediment export by the Senegal River became significant (Nicholson and Flohn, 1980; Koopmann, 1981; Gasse and van Campo, 1994; Gasse, 2000; Mulitza et al., 2008; Nizou et al., 2010). This change in the fluvial contribution between the HS1 and MH slices is most distinct at the location of GeoB 9510–1 (Figs. 1d and 2a). There, the fine sediment is trapped at the eastern (upper) flank of a sea-mount, acting as a down-slope transport barrier (Mulitza et al., 2006). However, if the grain-size distributions are being considered only (Fig. 2a), the change in fluvial contribution between HS1 and MH can only vaguely be detected at other locations of the transect. The change in the relative clay abundances seems to be more evident, as these values double or even triple from HS1 to MH (Fig. 2a; Tables 4 and 5). However, the increase in clay content might reflect a weakened MH dust flux in the 40–63 μm grain-size mode.

This ambiguity can be resolved by magnetics, as the presence of fluvial sediments during MH is clearly revealed by the magnetogranulometric parameters in Figs. 4a and 5a,b. Fluvial sediments contain a higher fine-grained magnetite content in comparison to < 10 μm fractions from the HS1 time slice. The concentration of fine-grained fluvial magnetite also decreases faster with shore-distance than the aeolian contributions, as expressed by the much steeper ARM/IRM vs. grain-size gradient in the 1.6–20 μm range (Fig. 4a). Fine-grained magnetite is also present in the HS1 grain-size fractions (Fig. 4a), but the variation with shore-distance (especially within the two aeolian modes at 10–20 and 40–63 μm) is much lower, pointing to the extensive dust transport by the Harmattan. In cross-plot Fig. 5c,d, magnetogranulometry and magnetomineralogy are combined. Here, the fluvial contribution during MH is also differentiated from the aeolian by a higher fine-grained magnetite proportion and by a lower haematite-magnetite proportion. This cross-plot visualises that the haematite-rich trend of HS1 does lack the fluvial component.

We argue that this fine fluvial magnetite is either of igneous origin from the nowadays semi-humid upper reaches of the Senegal River or pedogenically formed in the overlying soils, which are very common there (Lyons et al., 2010 and references therein). A biogenic origin for very fine-grained magnetite must always be considered as an alternative explanation for stronger sediment magnetic properties. Fossilised magnetosomes (very crystalline particles of 0.02–0.1 μm) created by magnetotactic bacteria are very common in marine surface sediments (e.g. Petersen et al., 1986; Just et al., 2012a). Additionally, Oldfield et al. (2009) showed by means of grain-size-based rock-magnetism on various terrigenous samples that magnetosomes tend to settle mainly in fractions of 2–4 μm. This was explained by their clustering to chains and aggregation to clay minerals. However, in a magnetic end-member study on sediment cores from the continental slope off the Gambia River, Just et al. (2012a) found that higher sedimentary magnetosomate concentrations were linked to higher MH riverine input of fresh terrestrial organic matter (Zarriess and Mackensen, 2010). Thus, even if the higher magnetisation with fine clastic grain-size is due to biogenic magnetite, the latter is valid to be used as indirect marker for stronger riverine input in our study area and fully supports our palaeoclimatic and transport-related interpretations.

Since budgeting of aeolian and fluvial sediment fluxes is beyond the scope of our study, we deal only with relative concentrations (Fig. 2a; Tables 4 and 5) and not with accumulation rates. A quantification of the fluvial sediment flux would require determination of sedimentation rates from more precise age models and a sediment volume calibration of magnetic markers as described by Just et al. (2012b). However, by combining magnetogranulometric (Figs. 4a and 5a,b) with magnetomineralogical (Figs. 4b and 5c,d) parameters from two contrasting climate situations, we seem to be able to distinguish between fluvial and aeolian sediments in the fine 1.6–10 μm grain-size range. This is interesting, as foregoing studies in this area could not untangle the “mixed” fluvi-aeolian <10 μm sediment fraction (e.g. Koopmann, 1981; Groussset et al., 1998; Itambi et al., 2009). These findings were possible primarily because of the additional gain of grain-size specific information, which goes further beyond those obtained from exclusive bulk investigations. In particular, the magnetogranulometric parameter ARM/IRM on a grain-size specific basis appears to be sensitive in detecting selective magnetite dissolution, while it also contributes to a more robust provenance study.

6. Conclusions

The simultaneous action of multiple environmental processes, ranging from sediment dynamics to climate-related weathering of source rocks, defines grain-size distributions of terrigenous marine sediments. Unravelling these processes from proxy records is a non-trivial task. This pilot study returned to the well-studied continental slope sediments off northern Senegal as an application study to fathom the potential of integrated grain-size and environmagnetic analyses. By applying rock-magnetic measurements not just to bulk samples, but to a wide range of well-defined grain-size fractions obtained by wet sieving and settling methods, many new interpretational options arise. Since this approach also distinctly enlarges the number of processed samples, it is not feasible to deploy a continuous high-resolution record. Therefore, we decided to focus on two contrasting Northwest African climate phases, the arid Heinrich Stadial 1 and the humid Mid Holocene.

The joint consideration of sedimentological and magnetogranulometric parameters appears to be a sensitive and powerful approach for a differentiated understanding of source and transport mechanisms. The insights would not be achievable solely from an interpretation of grain-size distributions and bulk magnetic measurements. Especially, the magnetogranulometric parameter ARM/IRM on a grain-size specific basis appears to be sensitive in detecting selective magnetic dissolution.

We departed from the assumption of two “aeolian” grain-size modes of 40–63 μm (north-easterly trade winds) and 10–20 μm (Harmattan) and were able to verify that these grain-size modes contain different proportions of fine (as inclusions) and coarse (separate) magnetite particles. The trade wind and Harmattan modes occur in the sediments with different haematite to magnetite proportions. Haematite occurs as fine-grained coatings on larger quartz particles and clay minerals. Beyond distinguishing transport systems, these magnetogranulometric “fingerprints” mirror the formation conditions of the magnetic particles at their respective sediment source (trade winds: Senegalese coastal dunes – Harmattan: southern Sahara and Sahel). Palaeoclimatic variations in aeolian transport dynamics could hence be identified by their peculiar grain-size modes. Changes in the palaeoclimatic conditions of the Sahel could be detected by changing proportions of haematite-coatings within the Harmattan sediment mode. Fluvial and aeolian sediments in the 1.6–10 μm grain-size range could be differentiated by the higher fine-grained magnetite content of the former. Fluvial fractions show a much greater shore-distance dependency as aeolian fractions.

Particles in the 63–500 μm range were identified as autochthonous benthic foraminifera and planktonic foraminifera settling from the surface waters above and probably being filled with fine magnetite-bearing sediment during burial. This was hinted at by the absence of consistent magnetogranulometric trends and the occasionally negative diamagnetic susceptibility values of these fractions.

We could show by comparing standard magnetic bulk-sediment measurements with those of individual grain-size fractions that it is not straightforward to tie bulk magnetic properties to certain grain-size fractions. This can become error-prone, particularly when sedimentological and environmagnetic investigations would be launched in areas, where no constraints are possible concerning the distribution, concentration and origin of magnetic minerals. In such cases, however, it would seem insightful to check also the grain-size dependency of magnetic properties, at least for a few selected test samples. Once the fractions
containing the magnetic mineral concentration-peaks are detected, the fractionation approach can be simplified and limited to just a few fractions of interest. Hence, the presented approach can improve the understanding of sediment transport processes, enable budgeting of sediment fluxes at complex depositional centres, facilitate provenance studies and thereby strengthen future palaeoecological and palaeoceanographic investigations.

Acknowledgements

The authors would like to thank the reviewers for their constructive comments which greatly improved the paper. We acknowledge Dr. A. Cletus Iambi for stimulating discussions and comments on the manuscript at an early stage. We thank the captain and crew of R/V METEOR cruise M65 who played a vital part in the recovery of the investigated sediment cores. Financial support for this study was provided to S.R. by the DFG through the European Graduate College “Proxies in Earth History” EUROPROX. This work contributes to MARUM Research Area Sediment Dynamics project SD2 – Climate control on large-scale sedimentary structures. Data presented in this study are available at the PANGAEA database (http://www.pangaea.de).

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