Do dune sands redden with age? The case of the northwestern Negev dunefield, Israel

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Abstract

The redness index of aeolian sand has been shown to be a promising qualitative spectroscopic method to define sand grain redness intensity, which reflects the extent of iron-oxide quartz grain coatings. This study investigates the relationship between redness intensity and optically stimulated luminescence (OSL) based depositional ages of sand samples taken from exposed and fully-drilled vegetated linear dunes in the northwestern Negev dunefield, Israel.

Sand redness intensity did not vary greatly along the Negev sand transport paths and dune sections dated to be active during the Late Pleistocene (~18–11.5 ka), Late Holocene, and modern times. No correlation was found between RI intensity (i.e., redness) and the depositional age of the sand. The relatively uniform RI values and sedimentological properties along most of the dunes suggest that sand grain coating development, and consequent rubification, have probably been minimal since the Late Pleistocene. Although it is possible that RI developed rapidly following deposition in a wetter Late Pleistocene climate, the drier and less stormy Holocene does not seem conducive to sand-grain rubification. Based on analyses of northern Sinai sand samples, remote sensing, and previous studies, we suggest that the attributes of the sand grain RI have been inherited from upwind sources. We propose that the sand grain coatings are early diagenetic features that have been similarly red since their suggested aeolian departure from the middle and upper Nile Delta.

1. Introduction

1.1. Sand color

Aeolian sediments in arid environments lack sedimentological characteristics such as organic remains, thus preventing palaeoenvironmental and palaeoclimatic reconstruction. Geochemical information, such as sand color intensities in quartzose sands, was often accepted as promising evidence for environmental reconstruction (Pye and Tsoar, 1987; Bullard and White, 2002).

Although it is poorly understood in terms of its formative process, sand color is a basic and easily described bulk property. Variation in sand redness intensity has been extensively described, mainly based on Munsell color criteria in arid tropical, humid tropical, and humid temperate climates across the globe, both in coastal (Lancaster, 1989; Ben-Dor et al., 2006) and inland desert dunes (such as Folk, 1976; Gardner and Pye, 1981; Anton and Ince, 1986). Whereas Munsell colors have been correlated to sand redness measured by field radiometry, the description of Munsell color is subjective, and radiometric measurements have higher precision (Bullard and White, 2002).

The reddish color of sands is understood to be the result of quartz grain staining, usually by thin orange to dark red coatings concentrated in grain pits and blemishes (Gardner and Pye, 1981; Hunt, 1991; Stanley and Chen, 1991; Besler, 2008). Scanning electron microscope (SEM) readings show that the surface of reddened quartz sand is covered in flakes and granular aggregates of hydrates of iron oxides, in which goethite (FeOOH) and hematite (Fe2O3) are the primary and secondary iron oxide compounds, respectively (Wopfner and Twidale, 1988; Pye and Tsoar, 2009). In time, these compounds fully coat the sand grain (Phener and Singer, 2001) in a process called rubification, which is defined as a change in soil color to yellow or red during intense weathering, thus liberating iron which then attaches to clay minerals (Mayhew and Penny, 1992). This quasi-pedogenic process involves the breakdown and weathering of iron-bearing minerals (Gardner and Pye, 1981) that usually originate from the parent rock (Folk, 1976; Anton and Ince, 1986) or in aeolian dust (Walker, 1979; Gardner and Pye, 1981; Hunt, 1991). Gardner and Pye (1981) and Anton and Ince (1986) hypothesized that sand grain redness is acquired following deposition without direct connection to the
parent rock in surface to near-surface oxidizing conditions in drained sand. Iron release and deposition is controlled by several environmental factors such as mineralogy, temperature, moisture, and water pH. When source factors and environmental conditions are homogenous, we can assume that varying hues of red in sand indicate different ages (Norris, 1969; Folk, 1976; Hagedorn et al., 1977; Walker, 1979; Gardner and Pye, 1981; Wopfner and Twidale, 1988; Goudie et al., 1993; White et al., 1997; Tsoar et al., 2008, 2009). Thus, in some cases, sand redness quantitation can potentially be a relative indicator of elapsed time.

There is considerable disagreement about the sand-grain rubification process (Besler, 2008). Time seems to be an important factor for both laboratory experiments of sand rubification and the natural rubification process, but there is no proof of a direct relationship between reddening and the age of sand using absolute dating. Although grain reddening has been simulated in the laboratory (Williams and Yaalon, 1977; Merrison et al., 2010), adapting this experimental data to natural processes is complicated. Grain residence time has been suggested as an important factor in reddening (Lancaster, 1989). Inland sand rubification is a slow process in arid and semi-arid climates, but in stable sand, distinguishable reddening can be attained in less than 10 k years (Gardner and Pye, 1981). The remotely sensed progressive rubification of Late Holocene Israeli coastal sands moving from the coast inland is suggested to be analogous to time (Ben-Dor et al., 2006), but this concept has not been proven by in situ dating for either inland or coastal dunes.

Previous studies have shown that most ergs, such as the Great Sand Sea in Egypt, the Taklamakan Sand Sea in China, Rub‘ al Khali in Arabia, and the Fachi-Bilma erg in the central-eastern part of the Tenéré Desert in Niger (after Besler, 2008), are homogenous in color. These studies, however, did not describe entire dune sections and neglected to include sufficient luminescence ages to investigate the relationship of sand redness to age.

1.2. Spectroscopy of sand redness

For Earth scientists studying aeolian processes, spectroscopic analysis techniques improve upon, and complement, the tools available for analyzing sediment properties. Various methods have been applied to spectrally measure sand redness. Free iron oxides that give soils their red color are identifiable across the VIS–NIR range and are spectrally active between 550–650 nm and 750–900 nm (Ben-Dor et al., 2006). Multi-spectral remote sensing and laboratory spectroscopy using different indices (Mathieu et al., 1998) have proven themselves reliable tools for quantifying sand redness, even though they are based solely on the visible red (R), green (G), and blue (B) bands (Madeira et al., 1997; White et al., 1997). Sand redness was evaluated using the R,G, B bands of a digital camera (Levin et al., 2005) and field spectroscopy was used to quantitate the iron oxide coatings of dune sand (Bullard and White, 2002). Iron oxide coated sands have also been multi-spectrally mapped based on laboratory measurements (Bullard and White, 2002; White et al., 2001, 2007). Recent remote sensing of central Saudi Arabian dune forms and sand redness demonstrate the complexities in understanding the significance of sand redness intensities (Bradley et al., 2011).

Redness indices are reliable in quantifying iron oxide sand-grain coatings. In the laboratory, spectroscopic spectral index measurements of sand grain coatings have been positively correlated with the Fe mass extracted from the grain-surface iron coatings using dithionite-citrate-bicarbonate (DCB) (Bullard and White, 2002; Ben-Dor et al., 2006; White et al., 2007; Tsoar et al., 2008).

Laboratory spectroscopy provides a uniform measuring environment without the physical and spectral constraints, such as changing surface cover (mixed pixels), variations in radiance relative to slope, atmospheric conditions, corrections, and varying observation angles, of remote sensing and field spectroscopy. Furthermore, remote sensing and field spectroscopy only measure the surface of the Earth (which can be covered by vegetation and crust), while laboratory spectroscopy can also measure sediment extracted from the subsurface.

Recent improvements in dune-drilling techniques (Stone and Thomas, 2008; Roskin et al., 2011a; Munyikwa et al., 2011) enable full dune profiles to be sampled. Hand drills can easily penetrate over 10 m into a dune and retrieve sand samples while preserving the dune's stratigraphy. Advances in the optical dating of quartz as the single aliquot regenerative (SAR) dose protocol (Murray and Wintle, 2000) have triggered a growth in the number of sand samples dated per study, significantly improving the chronological framework of studies. This study combines these improvements, which greatly facilitated the spatial and vertical quantitative analyses of dune sand rubification over time.

2. Study goals

In this study we challenge the hypothesis that dune sands redden with time. Our goal is to investigate the redness of the Negev and Sinai aeolian sands. We examine the relationship between the OSL age and redness intensity of dune sand sampled from the northwestern (NW) Negev dunefield (Fig. 1). If indeed dune sands do redden with time, deeper, more mature sands should be redder, and downwind dunes that have undergone longer periods of transport may also be redder.

By concurrently using spectroscopic measurements of sand grain redness acquired in this study and OSL ages of NW Negev dune profiles from Roskin et al. (2011a), we analyzed post-depositional changes in redness in situ and along Negev transport paths. Using the northern Sinai sand sample and multi-spectral remotely sensed sand redness values, we analyzed spatial trends in an effort to understand the transport, source, and formational controls of red coatings of sand.

3. Study area

The study area consists of the northern Sinai-NW Negev erg (Sinai-Negev erg), which is geopolitically split by the Egypt-Israel border (Fig. 1a). Unfortunately, the presently restricted access to the Egyptian portion of the erg precludes sand sampling in the northern Sinai Peninsula, making remote sensing a major tool for extracting data on Sinai sand properties. The erg lies to the north and downslope of a series of Mesozoic mountain ridges and Eocene highlands of mainly carbonate strata. Certain ridges, such as Gebel (Arabic: mountain) Maghara and Gebel Habil in the northern Sinai, contain erosional cirques and expose Jurassic-Lower Cretaceous Kurnub Group thick sandstone sections boasting colors of yellow, red, orange, and brown (Farag, 1955; Barakat, 1970) (Fig. 1).

The source of the northern Sinai dunes is believed to be the Nile Delta (Hunt, 1991; Tsoar, 1990; Roskin et al., 2011a), though this has not been proven. The coastal quartz sand dunes of the northern Sinai (Tsoar, 1976) and southern Israel (Ben-Dor et al., 2006) are whiter than those further inland. It is suggested that this is due to the bleaching, probably by the dissolution of oxides in water, of sand grain coatings in the submerged portion of the Nile Delta (Stanley and Chen, 1991) and of sand grains being carried by long-shore currents along and onto the northern Sinai and southern Israel coasts (Emery and Neve, 1960). Upper and middle Nile Delta quartz sand grains that have not been in contact with the coast are partially coated (Stanley and Chen, 1991). The Sinai sands east of the Delta mainly comprise active bare linear seif dunes (Tsoar, 1993; Misak and Draz, 1997; Abdel-Gailil et al., 2000; Rabbie et al., 2007).
Fig. 1. (a) A regional map of the Sinai–Negev erg. The erg, which stretches south and parallel to the southeastern Mediterranean coastline and extends eastward from the middle and upper northeastern Nile Delta, crosses the Egypt–Israel border (dotted black line) and extends into the northwestern (NW) Negev Desert. In northern Sinai, the mountain ridges of Gebels Maghara and Lagama block part of the dunes and expose Jurassic and Lower Cretaceous sandstones. The NW Negev dunefield was geomorphologically classified by Roskin et al. (2011a) into a northern (N on map), central, (C) and southern (S) dune encroachment corridors. A dashed black line distinguishes between the northern and central corridors while the Qeren ridge stands between the central and southern corridors. Gray box depicts Fig. 1b. (b) Dune axis mapping results, sampling site names and dune incursion/encroachment corridors (in capital letters). Dunefield regions [southwestern (SW), western and eastern] are also displayed and are referred to in the text.
that allow remote sensing of the surface sand properties. The northern Sinai dunes extend in a general west-east orientation into the NW Negev. Luminescence dating of the Sinai dunes has not been carried out.

The NW Negev dune field (N30°/E34°) constitutes the Israeli section of the Sinai—Negev erg, covering approximately 1300 km², of the Sinai—Negev erg (Fig. 1a and b). Its location at the downwind end of the erg, where sand has been deposited since the Late Pleistocene (Roskin et al., 2011a), is considered a suitable location to study sand rubification. The dunefield comprises stable vegetated linear dunes (VLD), whose vegetation cover, ranging from 5% to 15% (Siegal, 2009), provides minute organic material to the dune section (Blume et al., 1995). Similar to the linear dunes in the Sinai, the dunes elongated in a general west-east direction. The dune flanks are currently stabilized by biogenic crusts (Danin et al., 1989; Karnieli and Tsoar, 1995; Karnieli et al., 1996; Kidron et al., 2000). On the other side of the geopolitical border, the Sinai sands are barren of vegetation and biogenic crusts, and thus, they can be remotely imaged directly from space or air to measure sand redness. The southern dunefield corridor, OSL-dated to be slightly older than the central and northern dunefields (Roskin et al., 2011a), blocked and diverted ephemeral streams (wadis). Nahal Besor is the only such watercourse that transsects the dunefield.

The dunefield runs along a desert fringe between the climatic zones of the Mediterranean Levant and the global desert belts. Situated along the southern part of the wintertime cyclonic tracks of the Mediterranean Cyprus Low (a migratory surface low in the eastern part of the Mediterranean region accompanied by a cold air trough in the middle and high altitudes), the NW Negev dunefield receives approximately 150 mm of annual rainfall in the north and only 60–80 mm in the south. Accordingly, the biogenic crusts are several mm thicker to the north (Almog and Yair, 2007). Potential evaporation is 2000–2200 mm/year as measured at Nizzana in the southwest corner of the NW Negev dunefield (Fig. 1) (Stern et al., 1986). Further details of dunefield climate can be found in Littmann and Berkowicz (2008).

The Late Pleistocene climate along the Sinai—Negev erg has been interpreted to be stormier, wetter, and windier (Enzel et al., 2008). An archaeobotanical investigation of the Central Negev Highlands south of the Negev dunefield suggests a wetter Late Pleistocene between 18 and 10 ka (Baruch and Goring-Morris, 1997). During the Late Pleistocene until 14–13 ka, the northern Negev is suggested to have received 300–350 mm of rain (Vaks et al., 2006).

Previous spectroscopic and remote sensing research of the NW Negev that targeted surface sands and processes by applying different indices and complemented by laboratory work yielded important data but identified only general and undated spatial trends. Dunefield surface sand Musnells readings are 7.5–10 YR, value 4.4–7.5 and chroma 3–8 (Tsoar, 1976; Hunt, 1991; Blume et al., 1995, 2008; Campbell, 1999). These and other Musnells readings of Israeli arid and rubified soils have not been accurately converted to the redness ratio (after Mathieu et al., 1998) (Campbell, 1999) and spectral color ratios (Kelhamer, 2000), respectively.

Hunt (1991) suggested that the 2–4 μm thick amorphous iron coatings of Negev sand–grains occurred following the translocation of fines down the dune section (after Walker, 1979). Clays are retained on the grain surface as menisci films. Dissolution of fine fractions of heavy minerals contributes Fe leading to reddening of fines down the dune section (after Walker, 1979). Clays are retained on the grain surface as menisci films. Dissolution of fine fractions of heavy minerals contributes Fe leading to reddening of fines down the dune section (after Walker, 1979). Clays are retained on the grain surface as menisci films. Dissolution of fine fractions of heavy minerals contributes Fe leading to reddening of fines down the dune section (after Walker, 1979). Clays are retained on the grain surface as menisci films. Dissolution of fine fractions of heavy minerals contributes Fe leading to reddening of fines down the dune section (after Walker, 1979).

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4. Field and laboratory methods

4.1. Sampling methods

Sites in the Negev dunefield were selected at the western and eastern extents of each encroachment corridor to measure the hypothesized dune age-controlled color change from west to east (Fig. 1). Sand was sampled from full vertical sections of dunes and sand deposits in the Negev (Roskin et al., 2011a). This sampling strategy is hypothesized to account for an understanding of sand rubification trends, both spatially along sand transport corridors and temporally due to in situ changes in moisture content, illuviation, and other pedogenic processes. Most of the sections are exposed, which improved the reliability of the chronostratigraphic layers accumulated. The main dune encroachment occurred between 18 and 11.5 ka (Roskin et al., 2011a), and thick sand sections developed in the western dunefield. It has been suggested that dune elongation occurred in a windy climate during the Heinrich 1 and Younger Dryas cold events and that the eastern dunefield developed mainly in the Younger Dryas (Roskin et al., 2011b). Additional incursions and remobilizations have been dated to the Late Holocene (~2–0.8 ka) and modern times (150–10 years), respectively (Roskin et al., 2011a). Beyond these episodic events, the dunes were usually quasi-stable and probably partially encrusted (Roskin et al., 2011a). Observations and experimental results suggest that these relatively prolonged periods of dune quiescence may have enabled sand grain rubification.
Northern Sinai sediment samples, which helped validate the remotely sensed results, were acquired courtesy of Dr. Amihai Sneh, who sampled them for the Geological Survey of Israel during the late 1970s. To assess whether the southern Negev dunefield sands originated from a northern Sinai Late Cretaceous and Jurassic source, samples were taken from Negev analogs of these sands. Exposures of the Lower Cretaceous Kurnub Group (Hatira Formation) (Nubian) sandstones and Jurassic Inmar Formation sandstones in the Ramon erosional cirque in the central Negev Highlands, 30 km south of the dunefield, were sampled (Fig. 1a). Several samples were collected near representative sampling sites of Wenkert (2006) and Tsoar et al. (2008) and several samples from throughout the Negev dunefield, measured by Wenkert (2006) and Tsoar et al. (2008) were re-measured to ensure that our RI measurements were consistent with the previous work.

4.2. Spectroscopic measurements and indices

Laboratory spectroscopic preparation included carefully measuring 60 cc of split loose sand that was room-dried at 20 °C for 24 h in plastic plates to evaporate water vapor and eliminate condensation during measurement. To preserve the components that give the sample its natural color, samples were neither sieved nor purified. Sand samples were gently hand-ground to break up pedons. Immediately prior to measurement, the sand samples was transferred to a 4 × 4-cm opaque plastic black box and gently shifted to create a flat surface. Sand reflectance was measured using a contact probe of an ASD (Analytical Spectral Device) Fieldspec spectrometer (covering the VIS–NIR–SWIR spectrum (350–2500 nm) with an electrically-powered built-in Tungsten (1000 W) lamp at 45°. The contact probe was placed in a specially prepared wooden probe muzzle designed to ensure a uniform measurement distance of 1 cm of the probe edge from the sand surface. Four measurements were taken (every 90°) for each sample to take account of the Bidirectional Reflectance Distribution Function (BRDF).

All readings for each sample were averaged. The spectral bias between internal sensors at around 1000 and 1800 nm was corrected and the redness index was calculated using Ben-Gurion University of the Negev’s Earth and Planetary Imaging Facility (EPIF) bias correction Matlab algorithm.

4.3. Spectroscopic indices

The dimensionless redness index \[RI = R^2/(B + G^2)\] was found to be a favorable index amongst the Mathieu et al. (1998) color indices for the quantitative spectral measurement of sand rubification (Ben-Dor et al., 2006; Levin et al., 2007). RI values correlated to sand grain extractable iron oxide after Ben-Dor et al. (2006) \((R^2 = 0.89)\) for Israeli coastal dunes and after Tsoar et al. (2008) \((R^2 = 0.67)\) for NW Negev dunes, suggesting compatibility of the index for quantifying sand grain coating redness. The RI was calculated using specific though different R, G and B bands by Ben-Dor et al. (2006) and Tsoar et al. (2008) (Table 1).

Our RI results using the Ben-Dor et al. (2006) and Tsoar et al. (2008) bands are positively correlated \((R^2 = 0.94)\). However, we chose the specific R, G, and B bands after Ben-Dor et al. (2006) due to their better correlation with extracted iron oxides.

4.4. Landsat imagery

Landsat 5 TM images (row 175, images 38, 39) from June 1987 (30 m/pixel) were used. Since 1982, the relatively bare Negev dunes have been closed to Sinai Bedouin livestock grazing and wood gathering, immediately leading to the rehabilitation of biogenic crusts and vegetation (Meir and Tsoar, 1996; Karnieli and Tsoar, 1995; Tsoar, 2008; Tsoar et al., 2008). By 1984 when the first Landsat TM images were taken for this area, developing Negev dune vegetation and crust covers (Karnieli, 2011) already created a bias such as shown in the Wenkert (2006) ferric index analysis based on Landsat imagery. The June 1987 dates were chosen for analysis of the sparsely vegetated Sinai sands due to their quality, cover, and time of year when annual plants that in have wilted. Another image taken in August 2003 was examined for control.

The images were corrected using an improved dark object subtraction method, assuming 1% surface reflectance for the dark objects (Chavez, 1996; Song et al., 2001) (Supplementary material A2). The normalized difference vegetation index (NDVI), run in order to map unvegetated dunes designated the Sinai sands applicable for measuring the RI.

To fit the single band ASD Fieldspec spectrometer-measured RI to the RI of wide-band Landsat multispectral reflectance, the ASD Fieldspec spectrometer RI values were recalculated by resampling to match the reflectance spectra to Landsat spectral resolution (Supplementary material A3). An \(R^2\) correlation of 0.90 was found between the ASD Fieldspec spectrometer RI and the resampled bands (Table 1). Regional redness index maps of northeastern Sinai and of the NW Negev sands were processed using the RGB bands (Table 1).

4.5. OSL dating laboratory procedures

OSL dating used a modified SAR dose protocol to measure the equivalent doses. Purified quartz sand-grain fractions of 125–177 μm were measured on Riso TL/OSL readers. Gamma and cosmic dose rates were mainly estimated from burial depths. α and β dose rates were calculated from concentrations of the radioactive elements (K, Th and U) in the sediments measured by inductively coupled plasma atomic emission spectrometry. Further details and discussion regarding the accuracy and reliability of the OSL results are described in Roskin et al. (2011a).

4.6. Sedimentology

To better understand other sedimentological factors that may promote or hinder sand-coating redness, we investigated the link

### Table 1

Redness index (RI) band data from previous studies and the Landsat TM images (http://landsat.usgs.gov/about_landsat5.php), and their inter-relationship and relationship to Fe mass of sand-grain coatings.

<table>
<thead>
<tr>
<th>Source</th>
<th>R-Band (nm)</th>
<th>G-Band (nm)</th>
<th>B-Band (nm)</th>
<th>(R^2) of RI vs Fe mass extracted by dithionite-citrate-bicarbonate (DCB)</th>
<th>(R^2)</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tsoar et al. (2008)</td>
<td>640</td>
<td>510</td>
<td>460</td>
<td>0.67</td>
<td>0.94</td>
<td>between Tsoar et al. (2008) and Ben-Dor et al. (2006) R, G, B bands</td>
</tr>
<tr>
<td>Ben-Dor et al. (2006)</td>
<td>693</td>
<td>556</td>
<td>477</td>
<td>0.89</td>
<td></td>
<td>After Mathieu et al. (1998)</td>
</tr>
<tr>
<td>Landsat TM images</td>
<td></td>
<td></td>
<td></td>
<td>0.90 between ASD Fieldspec spectrometer derived RI to RI by resampling the ASD spectra to Landsat bands</td>
<td></td>
<td></td>
</tr>
<tr>
<td>(row 175, images 38, 39)</td>
<td>Band 3: 630-690</td>
<td>Band 2: 520-600</td>
<td>Band 1: 450-520</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

between sand redness and particle size distribution (PSD). PSD was measured mainly to investigate the contribution of the sand-silt ratio to rubification and to establish a cut-off range for defining “sand” samples. This was carried out by laser diffraction (using a Malvern Mastersizer MS-2000). Samples were split into 5-g portions, sieved to <2 mm, and stirred for dispersion for 10 min in sodium hexametaphosphate solution followed by ultrasonification for 30 s. Three replicate aliquots, later modified to two aliquots due to good reproducibility of the results, were run for each sample. Each aliquot was subjected to three consecutive 5-s runs at a pump speed of 1800 RPM. The raw laser diffraction values were transformed into PSD using the Mie scattering model.

Sand samples from representative dune and interdune sections were measured for moisture content by oven-drying.

5. Results and discussion

5.1. Redness index properties

The reflectance spectra of seven samples from sections throughout the Negev dunefield at depths ranging from 2 to 10 m are similar with respect to spectral features, although slight differences can be noted in the average reflectance and brightness (Fig. 2). RI and sedimentological results are presented in Supplementary material A1. The re-measured samples of Wenkart (2006) and Tsoar et al. (2008) had an $R^2 = 0.88$ correlation with the RI results of this study and samples collected near Wenkart (2006) sites displayed correlative RI values. This indicates that our RI measurements were in accordance with, and comparable to those of previous measurements.

Northwestern Negev dunefield RI values ranged from 21 to 87. Sand samples of both dune and interdune sand sections displayed relatively similar sand color, as observed in the sampling plates, and corresponding RI down their vertical profile (Figs. 3 and 4; Supplementary material A1). To obtain a representative RI value for each section in order to map general spatial trends, an average RI value and standard deviation of each section was calculated. Standard deviations for the RI of sections did not usually exceed ~15% (Supplementary material A1; Fig. 5).

The RI values of the Negev sands do not correlate with their OSL age (Fig. 6). In each section, OSL ages are naturally more mature with depth (Supplementary material A1; i.e., Fig. 3). At a regional level, due to variances in sand sedimentation rates in the Negev dunefield, OSL ages cannot be tied to specific depths. RI values per each section do not necessarily intensify with depth (Figs. 3 and 4).

Spatial changes in the RI show that the southernmost encroachment corridor is significantly redder than the central corridor, which is the least red (Fig. 5), in agreement with changes identified by Tsoar et al. (2008) for surface sands only. However, eastern sections and, in some cases, even more in their lower sections (i.e., Retamim ID; Fig. 1b; section 14 in Fig. 5b; Supplementary material A1), are also slightly redder (Roskin et al., 2011b). An outstanding section is Baladiya (Fig. 1b; section 4 in Fig. 5b), in the northeast corner of the dunefield, which shows the highest RI values in the entire NW Negev dunefield (Supplementary material A1; Fig. 5).

Compared with the RIs of Negev sands, those of the Ramon Cirque Lower Cretaceous sands proved significantly variable (RI = 7–98), with red and purple sands exhibiting higher RIs. Variability was also found in the light-brown to brown Jurassic sands (RI = 31–137) (Supplementary material A1).

5.2. Sedimentology and RI

The Negev sands are mainly fine-grained (125–250 μm). All sands were found to exhibit unimodal distributions, usually around 150–220 μm. Fine (silt and clay) content of dune sand is usually less than 20% and per section the fine values are often quite uniform. Samples with over 30% fines were not included in the RI analysis. Interdune silty sand units that interchanged with fluvial sourced loams and some dune bases had up to 30% fines (16 samples). Sand RI usually did not positively correlate with fine content. Sand moisture and XRD-derived sand mineralogy did not correlate to RI. Sand moisture was very low, between 0% and 2%. Sections from different parts of the dunefield have different moisture profiles. The relative abundance of calcite to plagioclase and quartz in several samples, mainly from the fringes of the dunefield (Roskin et al., 2011a), did not show a positive link with RI values. This may be spectrally explained by the absence of calcite absorption in the measured RI bands and exemplifies that while post-depositional mineralogical mixing is probably occurring, it does not affect sand-grain coatings and color, signifying that sand-grain redness has probably been inherited at least since deposition.

5.3. Sinai sand data

Except for local variability due to littoral and fluvial processes, the northern Sinai sand samples, at first glance, appear slightly less red than the Negev samples. The highest Sinai RI value is 44 while...
several Negev RI values top 80. The lowest Sinai sample had an RI of 18, similar to the minimal RI values (DF 690; RI = 21; Fig. 1b; section 6 in Fig. 5b; Supplementary material A1) of the Negev sand. The RI values of the Sinai samples showed spatial variability, although the limited and sporadic amount of data precludes the identification of spatial trends and does not allow a consistent comparison with the Negev RI values (Fig. 5). The lowest RI (=18) was recorded for a coastal sample, as these sands were probably bleached during their seaward transport path (Emery and Neev, 1960; Ben-Dor et al., 2006). However, low-RI (=18) samples, where also retrieved at Gebel Libni in the south and Wadi Khareidin (RI = 21) located approximately 30 km west of the Egypt–Israel border were higher RI values are identified by remote sensing and to the east for the southern corridor in the Negev (Fig. 5). These sands may be fluvial or lacustrine reworked sand (Sneh, 1983; Kusky and El-Baz, 2000; Roskin et al., 2011c) that has undergone abrasion and/or bleaching. Thus, most of the samples are comparable and exhibited RI values from 24 to 44 RI, similar to the range found in the NW Negev.

Satellite imagery can help us overcome the lack of samples for extracting spatially continuous RI for the Sinai sand surface. Where NDVI values were below 0.09, representing active dunes (Youhao et al., 2007) RI maps for the Sinai sands derived from the 1987 Landsat image show values of 20–40 RI that are similar to the spectroscopically measured values of the Sinai sand samples (Fig. 5; Supplementary material A1). The RI map also shows limited change in RI across the central Sinai sand body that continues into the central corridor of the NW Negev dunefield.

6. Discussion

6.1. Controls of in situ sand rubification

The similar RI values along the Negev dune and sand sections for sands of different ages are quite striking. They indicate that these sands are probably not subject to rubification processes, but instead, to rather limited and weak pedogenic alterations (Blume et al., 1995), and therefore, these sections have been in a fairly steady state since their deposition. Similarly, dune section sands in the southern Kalahari exhibit a homogeneous red color (Stone and Thomas, 2008) (Munsell 2.5 YR 4/8-R), and the uniform iron coatings are explained as pre-depositional features that have not undergone color change since deposition. Whereas water (moisture), dust (fines), and minerals have been suggested to contribute to sand rubification (Walker, 1979), we propose that the Negev sand sections have not undergone substantial leaching, infiltration of fines, and the dissolution of dust and heavy minerals to have generated subsequent oxidation and grain-coating growth.

The minute sand moisture variations of 0–2% are mainly controlled by crusts and seasonal sand movement of the upper dune section. Annual and seasonal rainfall infiltration studied in the southwest dunefield usually does not infiltrate the dune section to depths greater than 1–2 m (Yair, 2008). Beyond that depth, dune moisture is relatively steady and low. Gev (1997), who studied a dune east of Nahal Besor in the Negev dunefield (Figs. 1 and 5), also suggested that a limited amount of water percolates through the dunes and that the deep dune section usually has a steady moisture content. However, during the Late Pleistocene the northern Negev is suggested to have been rainier (Vaks et al., 2006) and to have experienced substantially higher loess-forming dustfall than today (Crouvi et al., 2008, 2009). In that rainier climate, rainwater probably often percolated and the dune section probably had higher moisture content.

Red colorization results from the presence of ferric oxide, which is derived from the weathering of iron-bearing minerals such as augite, olivine, hornblende, and epidote (Gardner and Pye, 1981) that are not abundant in Negev sands (after Hunt, 1991). Though the Negev sands are quartz-dominated (Hunt, 1991), we cannot prove that the sands did not previously contain mafic minerals and that those minerals did not already decompose in the past. Late Pleistocene water percolation and heavy mineral and dust mineral dissolution shortly after dune deposition may have formed a uniform color through the sand section. Ultra-natural rates of hot water circulation and leaching experiments by Williams and Yaalon (1977) proved that iron-rich heavy minerals, mainly hornblende, laterate and precipitate iron on surrounding quartz grains.
Sample DF 83 of the Haluzit 1 section (Fig. 1b; section 3 in Fig. 5b; Supplementary material A1) of a sandy palaeosol dune substrate dated to 106 ± 19 ka (Fig. 3), situated below a significant hiatus and presumably covered previously by sandy to silty palaeosols (Roskin et al., 2011a), has an RI value of 38 similar to the main dune section dating to 15 ka to the Holocene (Supplementary material A1). This finding presents strong evidence that RI values do not increase with time during the Late Pleistocene, and it does not support the possibility that earlier deposits were lighter-colored sands that underwent rapid reddening processes following deposition. These results indicate that Negev sand redness has not changed since the drier Holocene.

Hunt (1991) identified a slightly positive relationship only between the amount of fine-grained heavy mineral content and the grain coatings of Negev surface sands, weakly suggesting that solute Fe caused sand-grain coatings. From the current data, we cannot prove a relationship between heavy minerals and sand grain rubification in the Sinai and Negev. Current silt and clay fractions also do not contribute significant weathered iron oxides to the dune section. This may be due to the present and past high calcite...
and quartz contents of regional dust (Littmann, 1997; Crouvi et al., 2008) that lacks ferric materials.

The upper dune section, suggested to enable ferric precipitation due to oxidation (Gardner and Pye, 1981; Anton and Ince, 1986), lacked the moisture and dust, in both the encrusted and active dune surface scenarios, essential to this process. The Negev VLDs of today are characterized by biogenic crusts that trap fines and limit rainfall percolation (Kidron et al., 2000; Yair, 2008). Where these crusts are absent due to burial by sand or decimation by trampling, seasonal activation of the sand of the upper dune section occurs even despite rainfall-induced sand moisture (Allgaier, 2008). This sand reworking mechanism releases trapped fines, promotes water evaporation from the upper dune section, and keeps the sand column in a relatively well preserved state that

![Redness index maps of northern Sinai and the Negev.](image)

Fig. 5. Redness index maps of northern Sinai and the Negev. (a) Map of the remotely sensed RI ranges of the Sinai dunes derived from Landsat images taken in June, 1987. The colored boxes mark the location and spectrally measured RI of the Sinai sand samples. Note the lighter coastal dunes strip by square designating the (low RI) 18 value. The darker RI values south of these dunes are due to agriculture in sandy soils and those west of Al Arish are probably due to sabkha-induced microbial mats. East of the southwest corner of the Negev dune field in Egyptian Sinai, are dune sands, unfortunately inaccessible for sampling that reddens close to the border. This area, opposed to the northern erg fringes is not known to be significantly different than the rest of the dunes. The reddening seems to be solely of the sands, due to an unknown reason. (b) Map of the incursion corridors and average spectroscopic RI and standard deviation for the sampled dune and sand sections of the Negev. The Negev dune field section numbers (in gray) correspond to the data presented in Supplementary Material A1. (For interpretation of the references in color in this figure legend, the reader is referred to the web version of this article.)
limits the factors and materials contributing to pedogenetic processes.

A mechanism for trivalent to bivalent Fe iron reduction of sand grain coatings in the Negev and Sinai dune fields, such as the bleaching process occurring in anaerobic conditions of inundated sands between active parabolic coastal dunes in Brazil, is not likely (Levin et al., 2007; Tsoar et al., 2009) as sands in the southern dunefield, amidst units of standing-water deposits (Magaritz and Enzel, 1990; Roskin et al., 2011) are not less red. Various color degrees of sand samples along sections down to depths of 8 m of inland vegetated linear dunes in the Simpson Desert, Australia, have also been found to display similar ages (Nanson et al., 1992). As this may be due to later reworking, it also suggests that sand redness cannot be directly attributed to depositional age and that it may be attained either prior to and/or shortly following sand deposition.

The lack of supporting and convincing evidence that the NW Negev sands reddened in situ following deposition seems to suggest that the red color of the sand grains was inherited before their deposition in the Negev.

### 6.2. Spatial and vertical distribution of sand redness

It seems that there were two sand-color types that encroached into the Negev, with the sands that initially encroached being redder. The southern encroachment corridor sections and lower parts of some of the eastern sections have relatively higher RI values (~55–75) (Figs. 4 and 5; Supplementary material A1). Based on OSL dating, they appear to slightly predate the central corridor sands (Roskin et al., 2011a), as suggested by Tsoar et al. (2008). The southern corridor is more arid than the dunes to the north (Fig. 1), and this should imply that the rubification processes shown to be connected to rainfall moisture are less intense. This strengthens our argument that the Negev sands were probably not reddened or bleached during the more arid Holocene. Therefore, this may imply that these sands were redder than their counterparts in the central corridor already during their initial encroachment in the Late Pleistocene, which, in turn, may suggest that the redder sands have a different sand provenance or stratigraphic position than the lighter-colored sands.

Sands of the eastern dunefield sections with higher RI values probably reached the Negev at a similar time to the southern corridor sands and were re-transported further east or covered by lighter-colored sands during the main sand encroachment at 18–11.5 ka (Roskin et al., 2011a). The eastern dunefield sections include the Baladiya section (Fig. 1b; section 4 in Fig. 5b; Supplementary material A1) in the northeastern corner of the dunefield that is dated to the main dune encroachment (~15.9–13.7 ka) but that has the highest RI values (average of 73). Also in the eastern part of the central corridor, the basal sand of the Retamim interdune section (Fig. 1b; section 14 in Fig. 5b) dating to ~27–23 ka is also redder (RI = 43–53) than the overlaying 11-m thick interdune and dune sands (RI = 20–37). This also strengthens the notion that the initial Negev dune sand was possibly redder and not reddened shortly after deposition. Thick, fluvial, brownish-yellow sand units along Nahal Besor IRS dating back to ~20 ka may have originated in these sands (Greenbaum and Ben-David, 2001; Ben-David, 2003) in the central corridor, but were then partially washed out during seasonal flow in Nahal Besor. Further east, the slightly darker sands of the Ramat Beqa section (RI = 42–64) may also be remnants of early redder sand that mixed with the later sand of the main encroachment, a process that reset the luminescence ages.

Sand grain collision during downwind transport in the Mule-shoe dunes of the SW United States has been hypothesized to abrade grain coatings, explaining an observed downwind decrease in dune sand color (Muhs and Holliday, 2001). Abrading sand grains for up to 500 h in an aeolian abrasion chamber has led to a decrease in sand spectral redness (White and Bullard, 2009) in support of this hypothesis. Assuming that the Negev sands did not acquire their color shortly following deposition in the wetter Late Pleistocene, this trend is not observed for the NW Negev dunefield. The observation that linear dune sand has been found to be redder than transverse dunes, possibly due to their longer stabilization episodes and less abrasion (Livingstone and Warren, 1996), holds for the Negev regarding the former dune type that are usually stable since the Late Pleistocene (Roskin et al., 2011a). These are additional indicators of the negligible changes in sand properties and sand-grain coatings during transport.

### 6.3. Sinai sand redness

The general fit between spectroscopic RI values for Sinai sand samples and the multispectral RI mapping provide a reliable picture of the redness intensity of the Sinai sands. These values cover the whole of the main (central) and northern dune body between the ridges of Gebel Maghara and the Mediterranean coast (Fig. 5). The similarities of the upwind remotely-sensed Sinai and central and northern corridor Negev sand sample RI values, suggests that the sands from the western Sinai and throughout

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the Negev have relatively constant RI values. This strengthens the understanding that Negev sand redness was inherited from the Sinai. Similarly, like the Negev sand dunes, the Sinai dunes show no significant color changes downwind of their transport path. Accordingly, and with reference to Livingstone and Warren (1996), we also suggest that the sand grains have not been significantly abraded.

The redder values for sands in the Negev southern encroachment corridor also shows higher RI similarity to remotely sensed upwind Sinai surface sands (Fig. 5). Directly west of the border, the Sinai sands are observed to be even redder than further upwind. For the southern corridor, however, the upwind sedimentological setting in Sinai is diverse and was therefore examined for possibly accounting for the increased sand redness. In the Sinai, the corresponding upwind southern corridor dunes are usually not as thick as those upwind of the central Negev corridor (Abdel-Galil et al., 2000), a situation that may give the sand grains of the Sinai greater contact with the underlying carbonate substrate. These dunes also block ephemeral watercourses of fine grained carbonate sediments, and their interdunes are infilled with bright silts that form the top section of the Wadi Al-Arish floodplain (Sneh, 1983). This is probably due to standing-water deposits from Wadi Al-Arish (Kusky and El-Baz, 2000), which is blocked, probably by dune-damming (Roskin et al., 2011c). Extensive floodplains are situated further west around Gebel Libni (Kusky and El-Baz, 2000). The readily apparent and significant deposits of fine-grained carbonate sediment adjacent to redder sands recall the unexplained proposed connection between plays and sand rubification in the Great Sand Sea dunes in Egypt (Besler, 2008) while in the Balearic Islands of Spain, calcium carbonate content lowers the redness values of sandy palaeosols (Wagner et al., 2011).

According to Besler (2008), the red sand color of the Great Sand Sea may be inherited mainly from Lower Cretaceous sandstone formations. This may also be the case for Sinai. Therefore, Jurassic and Lower Cretaceous sandstones were investigated as possible, albeit partial, sources sufficient to intensify the bulk sand color of the Negev’s southern encroachment corridor. The eroded Jurassic and Lower Cretaceous Kurnub sandstone outcrops of Gebel Maghara are located upwind of the southern corridor and are more than 100 km closer to the Negev than the presumed Late Pleistocene middle-to upper Nile Delta sand source (Roskin et al., 2011a). Surface dune sands near the base of Gebel Maghara have been qualitatively described as yellow (Farag, 1955), whereas Gebel Maghara’s Jurassic sandstones are iron-oxide brown (Barakat, 2003; Enzel et al., 2008) and may even have had a different location, an upper Late Pleistocene iron-stained sand layer overlain by a silty clay layer radiocarbon dated to ~20–15 ka is rich in heavy minerals, notably hornblende (Coutellier and Stanley, 1987). These generally yellowish-brown clay layers display oxidized patches suggesting a connection between thick quartz sand sections and the heavy minerals exposed to changing and mainly stagnant aquatic environments that cause intermittent oxidation.

Exposed sand samples from the lower (northern) Nile Delta and lower Nile River, along with Late Pleistocene core sands from the central Delta, were classified petrographically according to sand grain content as transparent or either partially or fully stained with iron-oxides (Stanley and Chen, 1991). The Nile Delta sand grains from cores 43.7 to 2 m deep (Late Pleistocene–Holocene) and currently exposed sands were found to be yellow-brown and partially (~50%) stained. These sands are also often interspersed with marsh and swamp deposits (Friel and Stanley, 1987) that may have contributed ferric oxides and influenced rubification. Desert dune sands in the western Delta found to be 90% partially stained with iron-oxides may also be source sands for the Sinai erg. Other facies found mainly in the northern Nile Delta, protruding into and near the Mediterranean coast (lagoon, beach, transgressive, and near shore) to the northwest of the Sinai-Negev erg, show only ~20% staining (Stanley and Chen, 1991) and can possibly explain the light color of coastal dunes as found in northern Sinai (Fig. 5a) and in southern and central Israel.

Petrographic analysis of Negev sand grains shows they are also partially to fully coated/stained with the ferric oxides that give the sands their reddish color (Wenkart, 2006) similar to the color of upper and central Delta sands. SEM analyses of NW Negev sand grains also revealed extensive, finely disseminated, non-crystalline, ferric oxyhydroxides (Hunt, 1991).

Linear dunes in the upper Nile Delta are slowly advancing eastward toward the Sinai (Misik and Draz, 1997). As the general direction of the East Mediterranean region sand-transporting wind has not substantially changed since the Late Pleistocene period (Ben-David, 2003; Enzel et al., 2008) and may even have had a stronger west-east sand-transporting wind component in the past (Roskin et al., 2011b), desert, fluvial, and older sand deposits from the central Delta may have been transported eastward into the NW Sinai. At this time, between 30 and 11.5 ka the Nile Delta was an alluvial plain where sands were prone to aeolian erosion (Stanley...
and Warne, 1993). Accordingly, relatively unstained sands in the lower Delta along the coast were probably not transported into northern Sinai. Therefore, we suggest that the Sinai sands inherited their reddish ferric coatings from deltaic sands. Further sedimentological and chronological work is required, however, to prove the source of the erg sands and to differentiate between the darker and lighter red-colored sands found in the Negev.

7. Conclusions

The results of this paper diverge from its initial hypothesis, and it challenges the prevailing assumption that sand grain red color intensity derived from iron oxide films on quartz grains may be positively correlated to the depositional age of the sand. Based on full dune and interdune sand throughout the NW Negev dune-field, the spectrally-measured RI of the Negev sands is not positively connected to sand OSL depositional age. We cannot rule out the possibility that Negev sands that have been in situ since the Late Pleistocene may have undergone pedogenetic processes and rubification shortly after their deposition in a rainier Late Pleistocene climate, though there is no supporting evidence for this. Since the Holocene, sand color has not changed. The current Sinai sands have similar RI values to the sands of the Negev, suggesting that the iron-oxide coating of the sand grains is an earlier, diagenetic characteristic of the sands.

Late Pleistocene to current Nile Delta sand grain stain intensity and mineralogy values derived from previous works constitute supporting (though partial) evidence that Nile Delta sands may be the main, already-red source of sand for the Sinai-Negev erg.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.aeolia.2011.11.004

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